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DEPARTMENT OF MINES AND PETROLEUM RESOURCES
HON. D. L. BROTHERS, *Minister*　　K. B. BLAKEY, *Deputy Minister*

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GEOLOGY
# of the
PINE VALLEY

Mount Wabi to Solitude Mountain
Northeastern British Columbia

by

J. E. HUGHES

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1967**

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GEOLOGY OF THE PINE VALLEY
Mount Wabi to Solitude Mountain
Northeastern British Columbia

SUMMARY

1. The map-area covers the Pine Valley, in the Rocky Mountain Foothills of northeastern British Columbia, from latitude 55 degrees 30 minutes to latitude 55 degrees 45 minutes north.

2. The exposed rocks are Triassic, Jurassic, and Cretaceous in age. The succession (Ladinian to Cenomanian) is between 10,000 and 20,000 feet thick, and mostly of marine deposition. Most stratigraphic units thicken westward.

3. In the Triassic, the Grey Beds contain limestones, dolomites, siltstones, and sandstones; and the overlying Pardonet Formation, argillaceous limestones with fossil shell beds of *Halobia* and *Monotis*.

4. The Fernie Group of Jurassic age consists of: thin limestone, interbedded shales and siltstones with cherty banding, the Nordegg Beds; followed by the Middle Shales; and in the upper part, interbedded shales, siltstones, and sandstones, the Transition Beds, which mark the change to Beaudette deposition.

5. The Beaudette Group of late Jurassic to early Cretaceous age has three formations: the Monteith, thick sandstones mostly, and with quartzites in the upper third part; the Beattie Peaks, interbedded shales, siltstones, and sandstones; and the Monach Formation, sandstones, with or without quartzite beds at the top. Facies changes and incomplete outcrops make it advisable to map Beaudette strata as an undivided unit to the west.

6. Coal measures overlying Beaudette strata are described by the term Crassier Group. The group is a continuous sequence of deposits in the field. They have a complex lithology: shales, mudstones, siltstones, sandstones, grits, conglomerates, and coals, laid down in cyclic repetition. The sequence is undivided to the west. Eastward it can be divided into the Brenot, Dresser, and Gething Formations, according to their sand/shale ratios, the nature of the cyclothems, and the distribution of the coarser clastics.

7. The Fort St. John Group contains alternations of shale formations, the Moosebar in the lower part, the Hasler, and the Cruiser, with sandstone formations, the Commotion, and the Goodrich. These units are marine, except for thin non-marine beds and coal measures in the Commotion Formation.

8. The Dunvegan strata are mostly non-marine, and they form the youngest formation of the Cretaceous in the map-area.

9. The sedimentary rocks were deformed and uplifted in the Rocky Mountain orogeny, one phase of the Laramide revolution.

10. The Rocky Mountain Foothills in the Pine area are in two parts: the Inner (or Western) Foothills and the Outer (or Eastern) Foothills. The structures of the Inner and Outer Foothills represent different tectonic styles.

11. Close folding and thrusting prevailed in the Inner Foothills, and the deformation resembles that of the Rocky Mountains on the west.

12. The Outer Foothills contain long anticlines separated by wide synclines, and here the deformation was restricted to the anticlinal folding, and faulting

along the anticlines. The Outer Foothills belong to the orogenic foreland. Their structures can be ascribed to block faulting in the basement, though positive evidence is lacking.

13. Folding in the Foothills is of parallel type. The fold forms are concentric, or angular. The angular folds (that is, parallel folds of non-concentric form) include those of cuspate and lambdate forms, in the writer's classification. Two types of concentric folds are distinguished—those of low and high fold amplitude respectively.

14. Concentric anticlines of low fold amplitude occur in the Outer Foothills. The Commotion anticline is the only example that has been drilled to depth. Its surface and subsurface structures differ considerably, and they are separated by a low angle thrust, or décollement, or a zone of complex folding and shearing.

15. High-amplitude concentric anticlines have fold centres at shallow levels, and may be underlain by cuspate and angular folds, as well as décollements. Concentric folds of high amplitude, cuspate, and angular folds occur together in the Inner Foothills.

16. According to the writer's view, cuspate anticlines developed by replacing concentric anticlines during the folding compression. Lambdate folds are not specially related to concentric folds, and the few examples noted lie close to fault planes.

17. The structures of the Pine Valley area form part of a larger tectonic framework, which includes: the junction of the northern and southern parts of the Rocky Mountains; and in the Plains, the junctions of three tectonic units of the foreland, the Halfway block, the Fort St. John arch, and the Alberta syncline. The junction of tectonic units of the foreland now occupy the site of the Peace River embayment, a palæotectonic basin of differential subsidence and sedimentation lying transversely to the former miogeosyncline along the site of the Rocky Mountains and Foothills.

18. Observations, and analyses of structural geology presented here, are of use in exploring for petroleum and natural gas. Natural gas has been discovered in the Pine River anticline. Important reserves of coal occur in the Pine Valley.

CHAPTER I.—INTRODUCTION

LOCATION

The Pine Valley lies in the Peace River District of northeastern British Columbia, and crosses the Rocky Mountain Foothills between latitudes 55 degrees 30 minutes and 55 degrees 40 minutes north, from Solitude Mountain to Mount Wabi on the east. These localities define the map-area of the Pine Valley (in text), for which Figure 1 is the index map, and Figure 2 shows the geology.

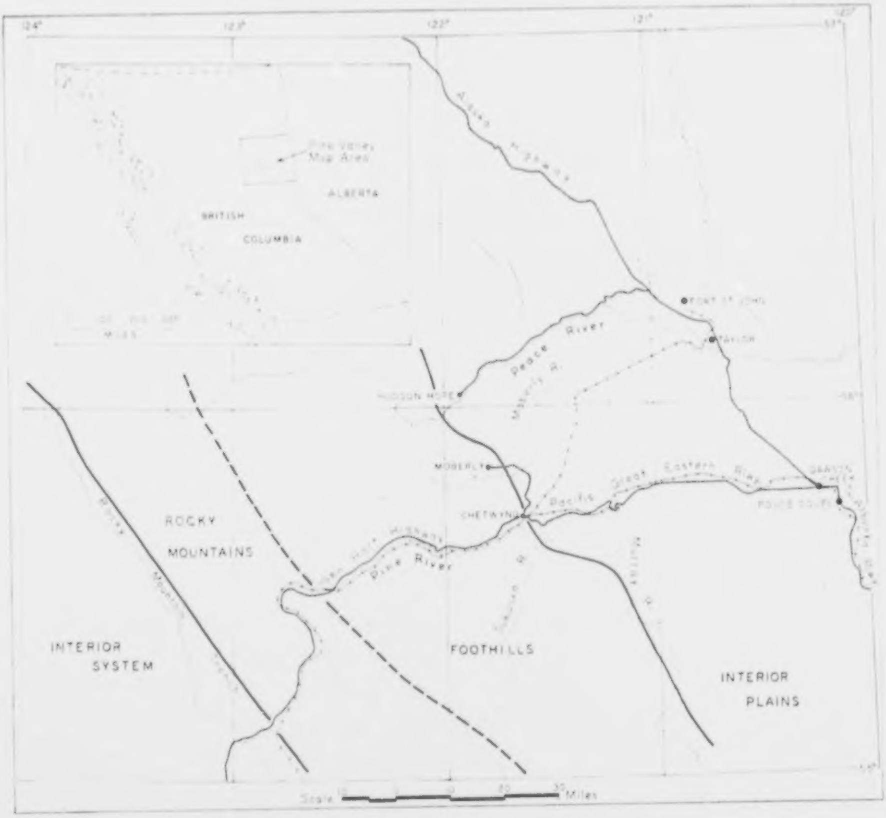
124°
123°
122°
121°
120°
57°
56°
55°
Pine Valley Map Area
ALBERTA
BRITISH COLUMBIA
Alaska Highway
FORT ST JOHN
TAYLOR
Peace River
Moberly R.
HUDSON HOPE
MOBERLY
CHETWYND
Pacific Great Eastern Rly
DAWSON CREEK
POUCE COUPE
Hart Highway
Pine River
Rocky Mountain Trench
ROCKY MOUNTAINS
INTERIOR SYSTEM
FOOTHILLS
INTERIOR PLAINS
Scale 10 0 10 20 30 Miles

Figure 1. Index map showing the Pine Valley.

FIELD WORK

The report is based on geological mapping in the Pine Valley (original scale, 2 inches to 1 mile) from 1954 to 1958, and a study of the Commotion structure in 1955 and 1956. Reconnaissance and stratigraphic work in the Foothills area of the Moberly and Peace Rivers (1955, 1959, and 1960) also contributed to this work.

ACKNOWLEDGMENTS

Dr. Stuart S. Holland and Mr. S. S. Cosburn introduced the writer to the Pine Valley, and Mr. N. D. McKechnie made available his notes and observations on the Willow and Hasler Creek areas. Field assistants were Y. Kawase and K. Newton, 1954; K. Falconer and C. Wright, 1955; Y. Kamachi, 1956; R. M. Luning, 1957; and A. Jellinek, 1958. The writer is grateful to them for their work, and to the residents of the Pine Valley for their interest and many courtesies.

Dr. J. A. Jeletzky identified the Upper Jurassic and Cretaceous faunas, and Dr. D. C. McGregor, the Cretaceous floras, at the Geological Survey of Canada, Ottawa.

Parts of the manuscript and its presentation were discussed by Drs. S. S. Holland and G. E. P. Eastwood. Full acknowledgments are due to Professors C. W. Stearn and P. Eakins of McGill University for their advice, reading, and supervision of part of this work as a thesis.

PREVIOUS GEOLOGICAL WORK

The first reconnaissances of the Pine Valley were made by G. M. Dawson (1881) and R. G. McConnell (1893). Later, geological work was directed to explorations for petroleum by J. C. Gwillim (1920), J. A. Dresser (1920, 1921), E. M. Spieker (1921, 1922), M. Y. Williams (1939); to explorations of coal deposits by J. Spivak (1944), W. H. Mathews (1947), N. D. McKechnie (1955); and to reconnaissance by M. Y. Williams and J. B. Bocock (1932), and C. R. Stelck (1942). The geology of parts of the Pine Valley, the Mount Hulcross-Commotion Creek area, was mapped by R. T. D. Wickenden and G. Shaw at a scale of 1 inch to 1 mile, and the Mount Bickford area by W. H. Mathews at a scale of 1 inch to 2 miles. The preliminary geological maps, Sheet 93-O Pine Pass by J. E. Muller (1961), and Sheet 93-P Dawson Creek by D. F. Stott (1961B), at scales of 1 inch to 4 miles, also include the Pine Valley. In stratigraphic work completed in the Pine area, G. E. G. Westermann (1962) zoned the *Monotis* beds of the Triassic; J. E. Hughes (1964) described Jurassic and Cretaceous strata of the Bullhead succession; R. T. D. Wickenden and G. Shaw (1943) classified the Fort St. John Group, and later D. F. Stott (1961C) redescribed the group. References in the bibliography mark the debt to F. H. McLearn (1918 to 1960) for his pioneer studies of Mesozoic fossils and stratigraphy in northeastern British Columbia. McLearn laid the basis for later geological work, and his studies apply to the geology of the Pine Valley.

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CHAPTER II.—PHYSICAL FEATURES

GEOGRAPHY

The map-area is part of the Rocky Mountain Foothills where they are crossed by the Pine River (Fig. 1). Here the Foothills are a terrain of mountain ranges and high plateau. They are mountainous in the west and merge with the Rocky Mountains without any topographic change. In the east they consist of a plateau which is partly crossed by several ranges of hills. Relief is considerable, 1,500 to 2,000 feet along the major valleys, and in addition 1,500 to 2,000 feet in the uplands above the valleys, or about 3,000 to 4,000 feet in all. The mountain summits in the Foothills have elevations of about 5,500 to 7,000 feet, similar to the Rocky Mountains. The climate is extreme. Winters are severe and the summers are cool and wet, yet changeable with some dry, warm spells. The area is thickly forested with spruce, pine, balsam, and poplar. Timberline is at 4,500 to 5,000 feet.

The Pine River rises in the Rocky Mountains and joins the Peace River in the Plains, and its valley served as an old route of travel across the Rocky Mountains. The Rocky Mountains and their Foothills are now crossed by the John Hart-Peace River Highway (No. 97) and the Pacific Great Eastern Railway. They follow the Pine Valley and the Pine Pass into the Missinchinka Valley, and link the interior and coastal parts of British Columbia with the Peace River District and northern Alberta, via the towns of Prince George and Dawson Creek. The Pine Valley is also followed by a pipe-line of the Westcoast Transmission Co. Ltd. carrying natural gas from the Peace River District to the southern part of British Columbia. A pipe-line for oil, laid by Western Pacific Products and Crude Oil Pipelines Ltd. on the same route, was filled in 1961.

The area is thinly settled, with most of the population concentrated in the eastern part, from Commotion Creek to Chetwynd. The economy is based on logging, ranching and farming, communications, and traffic. The valley floors have good soils. They are suitable for cultivation in the eastern part of the area, where the precipitation is moderate (around 30 inches). Large coal reserves have been proven in the Willow-Fisher Creek areas. Natural gas containing much hydrogen sulphide has been found in one well (Chapter V). Other resources include gravels, limestone, and building-stone.

PHYSIOGRAPHY

The Foothills border the Rocky Mountains and separate them from the Interior Plains to the east. Definition of the Foothills is partly a matter of geology and partly of physiography.

The Foothills of the Pine area are composite, and include two parts: a mountain belt in the west adjoining the Rocky Mountains, and a high plateau or a modified plateau on the east. Both parts were deformed and uplifted with the Rocky Mountains in the same period. The mountain-building is post-Cretaceous and referrable to the Laramide revolution. The Foothills are distinguished from the Rocky Mountains on the basis that Palæozoic rocks form the main exposures in the Mountains, and Mesozoic rocks outcrop in the Foothills. The Foothills contain a succession of strata that range from Triassic to Upper Cretaceous in age. The younger beds appear in order from west to east. The Cretaceous strata are continuous from the Foothills to the Plains in outcrop. Mass wastage and running water developed the initial pattern of topography. Post-orogenic and Recent uplifts accelerated these erosional processes. Glaciation modified the landforms and likely contributed to deepening the valley. Physiography and geology are related, and are best discussed in the two divisions of the Foothills which have been indicated. These are termed (i) the Inner or Western Foothills, and (ii) the Outer or Eastern Foothills. The divisions are made about the lines of Crassier and Willow Creeks.

(i) *The Inner or Western Foothills.*—The western part of the Foothills resembles the Rocky Mountains in physiography and landforms. It consists of long ranges of hills and mountains separated by longitudinal valleys aligned northwest-southeast, parallel with the geological trend. The ranges and their mountain summits have elevations between 5,500 and 7,000 feet. They have been eroded from resistant strata, limestones, sandstones, and quartzites, in thrust blocks and anticlines. The longitudinal valleys between the ranges lie at elevations of 2,500 to 4,500 feet. They are formed from downfaulted blocks, synclines, or complex folded structures. The valleys are heavily timbered and the hill and mountain ranges are bare or grass covered. The drainage pattern is trellis-form, the Pine forming the major transverse river. Its valley floor is broad and flat, with a cover of alluvium. The subsequent streams are incised to bedrock in the longitudinal valleys, and enter the main valley at grade or along steep courses. The Inner Foothills are a belt of strong folding and thrust faulting (Plate VI).

(ii) *The Outer or Eastern Foothills.*—In the eastern part, the close folding and faulting of the Inner Foothills give way to structures consisting of wide flat synclines separated by narrow faulted anticlines. The topography changes correspondingly. The Outer Foothills form a modified plateau which consists of long plateau belts separated by escarpments and cuestas trending north northwest-south southeast (Plate I). The plateau belts are developed on flat or slightly dipping strata in the synclines. The anticlines give rise to the escarpments and cuestas. These landforms are not obvious everywhere, and are not greatly differentiated in the plateau levels to the east.

In the map-area the plateau of the Outer Foothills lies between 3,000 and 4,500 feet. The general plateau level declines eastward, and becomes continuous with the High Plains (the Alberta Plateau). The boundary of the Outer Foothills is indistinct about the Pine Valley. It is drawn on geological grounds, and partly arbitrarily to include the more deformed structures within the Outer Foothills. The east front of the Chetwynd anticline makes a demarcation line. On the west margin of the Outer Foothills the plateau forms give way to a series of escarpments, developed from the anticlines forming the boundary of the Inner and Outer Foothills. In the Outer Foothills of the map-area the drainage pattern is composite, partly dendritic, partly radial, and partly trellis-form. The streams are incised and enter the main transverse valleys over steep courses or falls.

The main features of the landscape are the simple outlines of the plateau masses and the flat valley floor of the Pine, with its broad stretches, as much as 2 miles wide in places. Glacial meltwaters filled old lakes in the valley, and then drained away. Sediments of these old lakes made the flat valley floor, and were reworked by the Pine River and its tributaries. In Recent times, the river has lowered its course to about 400 feet below the floor of the old glacial lake from Browns Creek to Twidwell Bend. The downcutting is evenly distributed along the river's course through the Inner and Outer Foothills, along a fall of 2,275 to 1,850 feet elevation. A slight race of water in the gorge, 1 mile north of Twidwell Bend, marks where the river cuts through the glacial-lake sediments into bedrock.

The topography of the Outer Foothills is also related to their geology. About the Peace River, long ranges of hills separate the plateau belts, but disappear southward. They are formed from anticlines, and diminish in elevation as the anticlines plunge and terminate to the south-southeast. This plunge is a regional tectonic feature. The anticlines in the Outer Foothills area of the Peace River have high amplitudes, and the anticlinal ranges have high relief. The anticlines in the Pine Valley area are of low amplitude, do not form prominent hills or ranges, and have low escarpments along their faulted boundaries. The plateau belts of the broad synclines are undergoing stream dissection. Their sides along the Pine Valley are stepped; benches are formed by resistant sandstone beds, and shale formations are cut back.

GLACIATION

Erosional and Depositional Effects

The Cordilleran ice-sheet overrode the Foothills in Pleistocene time. It extended farther east, onto the west margin of the Plains, where it faced or adjoined the Continental (Keewatin) ice occupying the interior of North America. Later, when the climate ameliorated, deglaciation set in. The ice-sheets shrank, and their fronts separated and retreated. The intervening land was flooded by meltwaters and covered by a major lake. The lake bordered the east side of the Foothills. It was part of Lake Peace (of Taylor, 1960) at one stage until it was drained at the close of the Pleistocene ice age.

Ice covered most of the high ground in the Foothills. Glaciation developed cirques with tarn basins, as in the range between Little Boulder (Lillico-Marten*) and Big Boulder Creeks. They are present at altitudes of 5,000 feet in the Inner Foothills. Most of them face northeast, a result of geological structure, climatic factors, and the prevailing movement of the Cordilleran ice. Cirques are not prominent everywhere, and reduced, incipient, or decayed forms with low back walls are more common. Glacial erratics occur at altitudes up to 6,000 feet. Erratics of Triassic and Cretaceous rocks are abundant. Other erratics include Palæozoic limestones, quartzites, schists (including those of the Missinchinka Group), amphibolites, grey granites, and pegmatite-granites. The Cordilleran ice carried them from west to east in its advance. The distribution of glacial striæ and gouging indicates former ice movement at altitudes between 5,500 and 6,000 feet. In the Pine Valley, glacial striæ are aligned southwest-northeast near Narod Creek, and about this direction at Mount Wabi. The ice left boulder clays in the tributary valleys of the Pine, as in Fisher, Commotion, and Wildmare Creeks, and also on high ground between Moberly Lake and the Pine Valley.

The Pine Valley was deepened, either by accelerated erosion and downcutting in the late Tertiary or directly by glacial scour and sapping, or by a combination of these mechanisms. On the retreat of the Cordilleran ice, it was filled by a glacial lake which accumulated deposits of clay, silts, sands, and gravels. Marginal sands and gravels of the lacustrine beds can be traced to elevations of 2,450 feet in the Outer Foothills. The British Columbia Government Pine River No. 1 well cut 1,081 feet of unconsolidated fill. The bedrock of the preglacial valley floor now stands at an elevation of 930 feet or less near Commotion Creek (Appendix 3). The lacustrine beds are extensive, and they formed a continuous series along the Pine Valley in its course through the Outer Foothills, downstream from Fred Nelson Creek. Sands and silts predominate in the west; silts and clays occur in the east, clays and varved clays being found at lower stratigraphic levels (about 2,000 feet elevation and less, at Centurion Creek, and between Stone and Bissett Creeks). The lacustrine beds of this series are separate from minor deposits of gravels, sands, and silts, found in isolated outcrops near river level (elevations to 2,250 feet) from Crassier to Fisher Creeks, in the Inner Foothills. Sediments of an old glacial lake occur at higher levels, for McKechnie (1955) reported them at elevations above 3,700 feet about the headwaters of Willow Creek and at elevations of 3,000 feet along Hasler Creek. Lacustrine sediments are not distinguished farther west in the Inner Foothills. Here, sands and gravels deposited by the Pine River on its plain and the alluvial fans of its tributaries lie against bedrock along the valley walls. Depth of unconsolidated fill in the west part of the Pine Valley is appreciable, as indicated by well records: in Hunt Sands Sun Boulder b-74-D,* 1,460 feet of gravels and sands; in Triad Bush Mountain b-23-A (1), about 485 feet of gravels and sands (*see* Chapter V).

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* *See* review of geographic names in Note, page 131.

During deglaciation, the east front of the Cordilleran ice retreated from the Plains, back to the Foothills and the Rocky Mountains. The Cordilleran ice cover became restricted to high ground and to glaciers in the valleys. In the map-area, the front of the valley ice (glacier) retreated westward, and meltwaters formed a lake, which occupied parts of the Pine Valley to the east. There were two major still stands in the retreat of the valley ice. The first is shown by gravels and sands of the outwash about Jackfish Lake and the headwaters of Centurion Creek. This stand was probably a general feature along the border of the Foothills with the Plains: gravel and sand deposits fringe the north wall of the Pine Valley 3 miles downstream from Twidwell Bend, and some of them pass laterally eastward to silts with sands; patches of till and boulder clay follow the boundary of the Outer Foothills from Centurion Creek to the Moberly Valley; Moberly Lake is dammed by drift; the old course of the Peace River was seemingly blocked by a moraine across the Rocky Mountain Portage. A second stand in the retreat of the valley ice is shown by remnants of thick fluvioglacial gravels (elevations up to 2,450 feet), where Fred Nelson Creek enters the valley plain of the Pine. Seemingly, the second stand limited the glacial lake of the Pine Valley to the Outer Foothills at one stage of deglaciation.

Indistinct terraces marked only in discontinuous fringes of drift occur at elevations from 2,450 to about 2,650 feet along the north face of the Pine Valley, from Chetwynd to near Fred Nelson Creek. They may be old lateral moraines of the valley glacier or may record former strand lines of the valley lake in the Outer Foothills; a few cobble and boulder erratics are evident along their trace. The valley lake may have joined Lake Peace when it stood at 2,450 feet elevation or higher. Mathews (1963, p. 14) recognized the highest strand line of Lake Peace at 2,750 feet at present levels, near Fort St. John and Dawson Creek.

Lake levels were progressively lowered. Valley lakes of the Foothills emptied to the east, discharging into the proglacial-lake systems around the Continental ice-sheet. At one stage, water of the glacial lake occupying the Pine Valley in the Outer Foothills was impounded by the barrier of drift or ice, represented by gravels and sands, and pitted ground (elevation to 2,425 and 2,450 feet) about Jackfish Lake and the headwaters of Centurion Creek. The lake then drained eastward by a channelway at Jackfish Lake, and along an exit near Twidwell Bend.

Glacial-lake sediments of the Inner Foothills, from Crassier to Fisher Creeks, and the higher deposits about Willow and Hasler Creeks (McKechnie, 1955) belong to other phases of the deglaciation. At Falls Mountain, a dry gully (elevation 4,150 feet) cuts through the Commotion sandstones. It joins a channel, undercutting the boundary of the Moosebar shales and Commotion sandstones along the northeast face of the mountain. The cuts represent an overflow channel and spillway, part of the deglaciation drainage of the Willow and Falls Creek areas. There are few uncovered sections of glacial deposits in the Inner Foothills. Their sequence of deposition is not clear, but the following conclusions can be reached without conflict of evidence. The ice cover in the watersheds of Willow and Hasler Creeks, on ground at elevations from 3,000 to 4,000 feet, wasted and melted to form a temporary lake at high level, while valley ice in a thick glacier still filled the Pine Valley. (The Aletsch glacier damming the Märjelen See in the Alps is a modern counterpart.) The high escarpment of the Fort St. John beds and its remanent ice cover impounded the lake on the east, and isolated it from the contemporary(?) valley lake in the Outer Foothills. Finally, the retreat and decay of the Pine Valley glacier gave rise to ponds, or perhaps an extensive lake at low elevations (surface levels below about 2,450 feet) in the Inner and Outer Foothills or later in the Inner Foothills by itself. The sands and gravels between Crassier and Fisher Creeks belong to this stage of deglaciation. More fluvioglacial and lacustrine deposits may underlie the alluvium of the Pine Valley in the Inner Foothills.

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* Well names are given in this form, the word " well " not being incorporated in the name.

In Recent times the Pine River cut away and removed the upper deposits of the former glacial lake. The marginal deposits of the lake were left as terraced remnants along the valley sides, east of Fisher Creek. Sands excavated from one terraced remnant at 1,960 feet elevation, 1.5 to 2.5 miles west of Bissett Creek, contained small pelecypod and gasteropod shells.

DERANGEMENT OF DRAINAGE

Wastage and melting of the ice and the discharge of the glacial lakes was followed by resumption of normal drainage. The new drainage followed existing valleys, and its major pattern was determined by the pre-Pleistocene (late Tertiary) topography. It was superimposed on the remaining till cover and the glacial-lake floor of the Pine Valley. New courses were cut into drift, and into the underlying

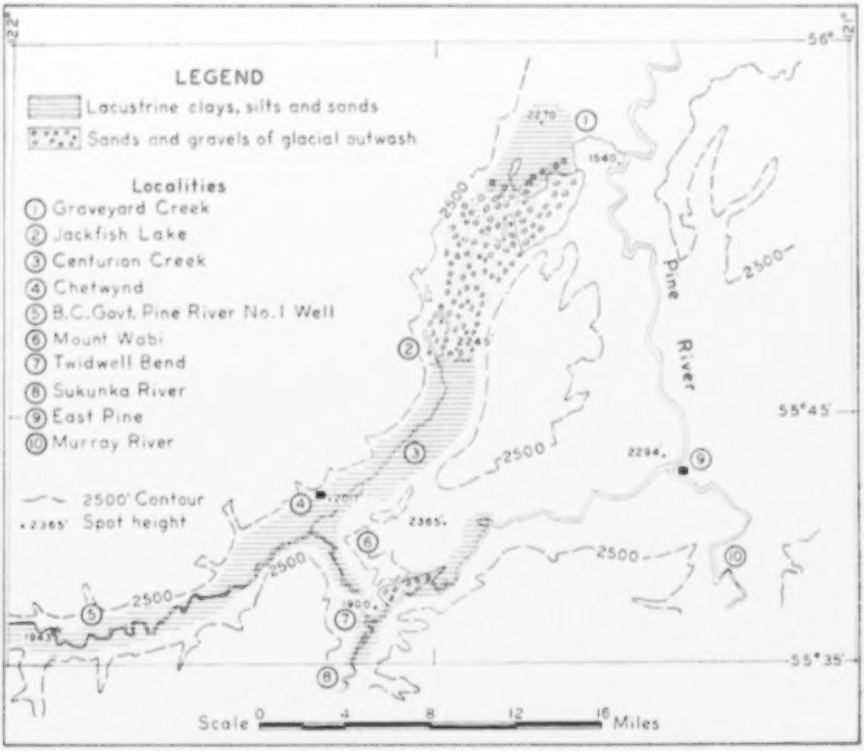
122°
121°
56°
55°45'
55°35'
LEGEND
Lacustrine clays, silts and sands
Sands and gravels of glacial outwash
Localities
① Graveyard Creek
② Jackfish Lake
③ Centurion Creek
④ Chetwynd
⑤ B.C.Govt. Pine River No. 1 Well
⑥ Mount Wabi
⑦ Twidwell Bend
⑧ Sukunka River
⑨ East Pine
⑩ Murray River
2500' Contour
2365' Spot height
Pine River
2500
2270
1540
2245
2294
2017
2365
1900
1943
Scale 0 4 8 12 16 Miles

Figure 5. Pleistocene deposits and the diversion of drainage, Chetwynd.

bedrock in places. Some diversion and interference of the drainage followed deglaciation—notably the deflection of the Pine River from a former north-northeast course downstream from Chetwynd.

The diversion of the Pine River after glaciation increased the length of its course by at least 15 miles. Near Chetwynd, near the east boundary of the Outer Foothills, the Pine alters its direction of flow, from northeast to southeast (Fig. 5). It flows southeast to Twidwell Bend, where it is joined by the Sukunka. The Pine River is incised 375 feet in glacial-lake sediments which underlie a valley plain at an elevation of 2,225 feet in this reach. In the lower part, 1 mile upstream from Twidwell Bend it cuts a gorge in bedrock. The Pine River resumes an east-northeast course downstream from Twidwell Bend. It is deeply entrenched in this lower reach to the junction of the Murray River, and the valley is steeply terraced. The terraces, many of which have slumped, were cut by the river in glacial-lake sediments, mostly clays and silts. In places the river cuts bedrock, as in the canyon noted by Dawson (1881, p. 42). Downstream from the entry of the Murray River, the Pine flows northward to the Peace River.

The preglacial course of the Pine River was northeast from Chetwynd. It lay in the broad valley of Centurion Creek, in which the present stream is underfit. Sands, silts, clays, and in places varved clays occupy the floor of the valley between elevations of 1,900 and 2,250 feet. The headwaters of Centurion Creek issue from pitted ground and deposits of fluvioglacial gravels. The gravels attain elevations between 2,425 and 2,450 feet. They present a steep ice contact face to the south, and abruptly block the valley of Centurion Creek. This barrier is breached by Jackfish Lake, elevation 2,310 feet.

To the northeast, the preglacial valley of the Pine is floored by a plain of gravels and sands, with silts bordering the gravels and sands farther northeast. The plain has an elevation of 2,400 to 2,250 feet, declining northward. It is cut by an entrenched river system, now abandoned, and marked by Jackfish Lake and other unnamed lakes to the north. A line of swamps along this channelway connects the lakes and then joins a diffuse and sluggish drainage which is taken up by Graveyard Creek. Graveyard Creek itself is underfit in the valley. It turns eastward and empties along a steep course into the present Pine River at an elevation of 1,540 feet.

As deglaciation continued, the glacial lake of the Pine Valley lowered in level and was impounded by the drift or ice barrier about Jackfish Lake and the headwaters of Centurion Creek. A spillway and outlet was obtained through Jackfish Lake (elevation 2,310 feet) and the series of deep channelways to the north. Finally the valley lake emptied to the south by an outlet at lower level, along the present course of the Pine River to Twidwell Bend, and downstream by other outlets along the course to the Murray River. This drainageway was then followed by the recent course of the Pine River, which later became entrenched in drift and bedrock.

Selwyn (1877, p. 54), in the first account of exploration on the Pine River, illustrated a view of the entrenched valley near Wartenbe Mountain. Dawson (1881, p. 41) noted the abandoned valley of the Pine along Centurion Creek, and the misfit of the Pine River in its present course from Chetwynd to Twidwell Bend. Again, Spieker (1922, p. 12) recognized the postglacial diversion of the Pine River. Both Spieker and Stott (1961B, p. 2) referred to the capture of the Pine River by the Sukunka. A hypothesis of river capture may explain the change of course of the Pine River since glaciation, but it discounts the evidence of glacial and fluvioglacial action which has been put forward.

The preglacial course of the Sukunka is less obvious; probably it lay along the present valley of the Pine downstream from Twidwell Bend.

CHAPTER III.—STRATIGRAPHY

INTRODUCTION

The strata outcropping in the Foothills area of the Pine Valley are of Mesozoic age, from Middle Triassic to Upper Cretaceous. The Mesozoic sediments of the Foothills were deposited on the shelf of a miogeosyncline. Sedimentary environments in the Foothills and the adjacent Plains were continuous and similar. In general, formations decrease in thickness from west to east across the Foothills and into the Plains. The Triassic strata are marine, and consist of limestones, calcareous shales, siltstones, calcareous and dolomitic siltstones, and sandstones. Marine shales predominate in the Jurassic. Sand deposition marked the close of the Jurassic and continued into the lower part of the Cretaceous. Marine and nonmarine sedimentation in the remainder of the Cretaceous left a thick sequence of sandstones, shales, coal measures, and deltaic beds. The Mesozoic sedimentation was interrupted by some marine recessions and advances, and breaks which were apparently not of long duration.

The sediments were folded, thrust faulted, and uplifted in post-Cretaceous times. The rocks are exposed in long fold and thrust belts trending northwest-southeast. The younger strata appear in the eastern part and the older strata in the western part of the Foothills. The succession is similar in the Foothills areas drained by the Peace, Moberly, and Pine Rivers, and most formations can be traced from one part to another in outcrop. The stratigraphy was first established by McLearn (1923, 1940) and Beach and Spivak (1944) in the Peace River area, by Wickenden and Shaw (1943) in the Moberly-Pine River areas, and by Mathews (1947) along the middle line of the Foothills between the Peace and Pine Rivers. Existing terms and classifications are followed here, except for the Bullhead strata (Hughes, 1964).

The stratigraphy is treated in two parts: firstly, the account of the outcrops; secondly, the account of drilled sections in the subsurface.

STRATIGRAPHY: OUTCROP

TRIASSIC

The Grey Beds and its formations, together with the Pardonet, compose the Triassic outcrops of the map-area.

The Triassic of northeastern British Columbia was studied by McLearn from 1930 onwards (*see* McLearn and Kindle, 1950; McLearn, 1960). His work established the succession and its faunas for outcrops in the Foothills. Hunt and Ratcliffe (1959) and, later, Armitage (1962) outlined the Triassic formations in the subsurface of the Plains. Colquhoun (1962) extended these formational terms to the Rocky Mountains and Foothills, a practice followed in part by Pelletier (1964). Thereby, the Halfway, Charlie Lake, and Baldonnel Formations of the Plains were equated with the Grey Beds. The Pardonet is now accepted as a formation after Hunt and Ratcliffe (1959), McLearn (1960), and Irish (1962). The term Schooler Creek Group (McLearn, 1921; Hunt and Ratcliffe, 1959; Armitage, 1962) can be taken to include the Grey Beds and its component formations, the Halfway, Charlie Lake, and Baldonnel, together with the Pardonet Formation.

Table I.—Table of Formations

SURFACE AND SUBSURFACE

<table>
<tr><th>Age</th><th>Group</th><th colspan="5">Stratigraphic Units: Thickness (in Feet): Lithology</th></tr>
<tr><td rowspan="2">QUATERNARY</td><td>Recent</td><td colspan="5">Alluvium.</td></tr>
<tr><td>Pleistocene</td><td colspan="5">Lacustrine clays, silts, sands. Fluvioglacial sands and gravels. Glacial till.</td></tr>
<tr><td rowspan="12">CRETACEOUS</td><td></td><td colspan="5">Dunvegan Formation (+<i>1,194</i>). Sandstones, shales, and siltstones; minor conglomerate; few thin coals; largely non-marine.</td></tr>
<tr><td rowspan="5">Fort St. John Group</td><td colspan="5">Cruiser Formation (450 to 850). Shales and mudstones; minor thin sandstones; marine.</td></tr>
<tr><td colspan="4">Goodrich Formation (500 to 1,225). Sandstones; minor conglomerate; some shales; marine.</td><td>* Shales with sandstones (<i>760</i>). Shales; four arenaceous members.</td></tr>
<tr><td colspan="5">Hasler Formation (<i>785</i> to 1,100). Shales with minor thin siltstones and sandstones; marine.</td></tr>
<tr><td colspan="5">Commotion Formation (<i>1,417</i> to 1,425). Marine sandstones, conglomerates, and shales; some non-marine beds; thin coal measures at top in outcrops.</td></tr>
<tr><td colspan="5">Moosebar Formation (<i>1,083</i> to 1,400). Mudstones and shales; minor sandstones; marine.</td></tr>
<tr><td rowspan="3">Crassier Group</td><td colspan="3" rowspan="3">Crassier Group Undivided (+2,500). Coal measures, and coal measures with much sandstone; few conglomerates.</td><td colspan="2">Gething Formation (<i>513</i> to 1,800). Coal measures.</td></tr>
<tr><td colspan="2">Dresser Formation (<i>670</i> to 1,200). Coal measures with thick sandstones, grits; some conglomerates.</td></tr>
<tr><td colspan="2">Brenot Formation (<i>305</i> to 750). Coal measures with thin coals and thin or barren cyclothems.</td></tr>
<tr><td></td><td colspan="4"></td><td>* Chetwynd Beds (<i>245</i>). Sandstones, quartzites, shales; thin coal measures.</td></tr>
<tr><td rowspan="3">Beaudette Group</td><td colspan="3" rowspan="3">Beaudette Group. Undivided (3,150). Mostly sandstones; minor quartzites; minor shales; marine.</td><td>Monach Formation (—300 to 400). Sandstones with or without quartzites in upper part; marine.</td><td rowspan="3">* Beaudette Group. Undivided (<i>688</i>). Sandstones, quartzites, siltstones, shales.</td></tr>
<tr><td>Beattie Peaks Formation (650 to 950). Shales, siltstones, sandstones; marine.</td></tr>
<tr><td rowspan="4">JURASSIC</td><td>Monteith Formation (+1,500.) Sandstones; quartzites in upper third; minor shales; marine.</td></tr>
<tr><td rowspan="3">Fernie Group</td><td colspan="5">Transition Beds (75 to ? 150). Shales, siltstones, sandstones; marine.</td></tr>
<tr><td colspan="5">Middle Shales (<i>313</i> to 600). Mostly dark-grey and black shales; marine.</td></tr>
<tr><td colspan="5">Nordegg Beds (50 to <i>97</i>). Limestones; followed by thin shales, siltstones, and sandstones; chert; marine.</td></tr>
<tr><td rowspan="4">TRIASSIC</td><td rowspan="4">Schooler Creek Group</td><td colspan="5">Pardonet Formation (? <i>nil</i> to 700). Argillaceous, silty limestones; aphanitic limestones; shell beds of <i>Halobia</i> and <i>Monotis</i>; marine.</td></tr>
<tr><td rowspan="3">Grey Beds</td><td rowspan="2">Division (ii)</td><td colspan="2">Baldonnel Formation (550). Limestones, with shelly fragmental and arenaceous limestones; marine.</td><td>* Baldonnel Formation (<i>425</i>). Dolomites; arenaceous dolomites; shales, siltstones, chert.</td></tr>
<tr><td colspan="2">Charlie Lake Formation (475). Limestones, dolomites, siltstones, sandstones, quartzites, anhydrites; marine.</td><td>* Charlie Lake Formation (+<i>163</i>). Dolomites; arenaceous dolomites; anhydrites; siltstones.</td></tr>
<tr><td>Division (i)</td><td colspan="2">Halfway Formation (400). Dolomitic and calcareous siltstones and sandstones grading to arenaceous dolomites; marine.</td><td>* Not drilled.</td></tr>
</table>

* Stratigraphic units and lithology belonging to subsurface section in east part of the map-area (drilled in Sun et al Chetwynd 14–20).

Italic figures denote thickness for stratigraphic units in subsurface, in east part of the map-area (drilled in Sun et al Chetwynd 14–20).

McLearn (McLearn and Kindle, 1950) gave the provisional names Dark Siltstones and Flagstones for older strata which underlie the Grey Beds. The two units, or equivalent strata, have since been brought under the formational terms Mount Wright (Colquhoun, 1962), Doig (Armitage, 1962), and included within the Liard Formation (Pelletier, 1964).

Schooler Creek Group

Grey Beds: Halfway, Charlie Lake, and Baldonnel Formations.—The Grey Beds contain clastic, evaporite, and carbonate sequences amounting to some 1,400 feet in thickness.

The term Grey Beds is retained here for surface mapping, as exposures are insufficient to define its formations in the field. A more complete record of the strata is available by combining mapping with subsurface sections drilled in Triad Bush Mountain b-23-A (1), and Triad B.P. Bush Mountain a-15-A. These show a threefold division of the Grey Beds: the Halfway, Charlie Lake, and Baldonnel Formations. The account of the Grey Beds is presented in two forms: firstly the field description, and secondly a description of its formations.

The Grey Beds have small outcrops east of Solitude Mountain in the Pine Valley. They are present along the valley sides and high ground to the north. The outcrops are forest covered, and have moderate relief, with resistant sandstones and dolomites making low and indistinct ridges above their surroundings. Exposures of the Grey Beds are discontinuous and reveal small intervals. In the less faulted ground, they are confined to the area between Silver Sands and Cairns Creek and to an old river bank near the Pacific Great Eastern Railway bridge across Mountain Creek. Mapping these exposures indicates two field divisions of the Grey Beds (Fig. 2).

Division (i): There are about 650 feet of beds in this lower field division. Its lower part is concealed and the base not proven. Division (i) contains calcareous and dolomitic sandstones weathering in buff and light colours, and associated thin, buff, and light-grey shales; dark-grey siltstones and fine-grained sandstones which are dolomitic and grade into arenaceous dolomites in places; some dark-grey shales in thin interbeds. A group of buff-weathering sandstones, in an exposure of 60 to 70 feet stratigraphic interval, marks the top of the division. The upper contact of the sandstones, unseen, is taken as the boundary with the upper field division of the Grey Beds.

There are few fossils in Division (i). Outcrops along the Hart Highway yielded *Lingula* sp. and obscure pelecypods.

Division (ii): The upper field division of the Grey Beds is 700 to 800 feet thick. The lower 200 to 250 feet are concealed. For the remainder, most of the beds are limestones, where seen. These include fine and coarse crystalline limestones and argillaceous limestones. Siltstone, calcareous siltstones, and sandstones, and some thin shales are lesser components. Much of Division (ii) is dark grey, but the crystalline, and shell fragmental and skeletal limestones, which characterize this division, have lighter grey colours. These limestones occur in association. The shell fragmental types contain pelecypod, gasteropod, and brachiopod remains, crinoid ossicles, and much sand residue after acid digestions. *Myophoria* sp. (of the Minetrigonid kind) and *Septocardia* (?*Pascoella*) sp. are notable in such associations, though they are patchily distributed; their shells are entire and fragmented, and are usually silicified. *Lima*? *poyana* McLearn is obtainable at one locality, the outcrop overlooking the Pacific Great Eastern Railway bridge at Mountain Creek. This fossil locality was reported by Williams and Bocock (1932), and its stratigraphic position is placed about 500 feet below the Grey Beds-Pardonet boundary.

Other collections contained *Lingula* sp. (probably *L. selwyni* Whiteaves), *Spiriferina* sp.; *Dielasma*? (*Terebratula*) cf. *julica* (Bittner); small brachiopods of terebratuloid appearance, and small rhynchonellids; cf. *Pteria? (A.) bittneri** (Böhm); *Modiolus* sp. (fragments); ?*Cassianella* sp. Strata of Divisions (i) and (ii) are separated by a recessed, concealed interval, which is included in the lower part of Division (ii) for mapping purposes.

The two field divisions of the Grey Beds are shown for the area northeast of Silver Sands Creek (Fig. 2). They also hold good on the southwest, in the fold and fault wedges under the Solitude thrust. The complex structures make it more practicable to treat the Grey Beds as an undivided unit here. Most of these westerly exposures belong to the lower division. The Grey Beds are close folded and faulted in the map-area. There is some calcite veining related to the deformation. Also, some beds are vuggy; one set of vugs, lined or filled with calcite, anhydrite, gypsum, sulphur, and with quartz as a late replacement, occurs in the dolomitic-clastic beds of the lower division; another set, lined with quartz and calcite, and flecked with pyrobitumen, occurs in limestones of the upper division. Fluorite is sparingly dispersed in the limestones. Calcareous boxworks and breccias of physico-chemical origin occur in the Grey Beds near the footwall of the Solitude thrust, 750 feet southwest of the West Pine Bridge along the Hart Highway.

The full succession of the Grey Beds, in surface and subsurface sections, can be presented as follows.

Halfway Formation.—The term Halfway denotes the lowest formation of the Grey Beds. It is mostly a carbonate-clastic sequence, containing dolomitic siltstones and sandstones, and arenaceous dolomites of similar appearance; dolomites and anhydritic dolomites, with anhydrite in drill cuttings; calcareous siltstones and sandstones; and few quartzites. Throughout the formation the grey and dark-grey colours of the dolomites and dolomitic clastics predominate over the calcareous clastics, which are light coloured. Dark-grey dolomites in the upper part of the formation are commonly veined and mottled with anhydrite and are associated with free anhydrite in drill cuttings. The origin of the latter, either as veins, intergrowths, vug fillings, or as bedded anhydrites, seems uncertain. Anhydritic material constitutes a subordinate proportion of drill cuttings from the upper third of the formation. The sand fractions of the Halfway are very fine to fine grained, resembling those of the overlying Charlie Lake Formation, in being well rounded, and quartzose, with minor feldspars. The Halfway Formation also contains a few thin interbeds of dark-grey shales and siltstones, which are argillaceous and variably calcareous.

The Halfway Formation overlies a sequence of dark-grey argillaceous to calcareous siltstones, which are referred to the Dark Siltstones unit of McLearn (McLearn and Kindle, 1950). A brief change of lithology marks the common boundary, but a transition is implied by similar siltstones in the Halfway, and the downward extent of light-coloured calcareous clastics as thin interbeds in the upper part of the Dark Siltstones. Another transition relates the Halfway to the overlying Charlie Lake Formation, leaving this boundary to be defined by arbitrary means. There are several indices for drawing the boundary: the distribution of light-coloured carbonates; the distribution of dark-grey dolomites and anhydritic dolomites; that of anhydrite itself; and that of clastics with carbonate and quartzitic matrices. These lithologies occur together and are repeated within the transition. By the first index, light-coloured limestones and dolomites are excluded from the Halfway Formation. They are characteristic of the Charlie Lake beds, and are a feature of clastic-carbonate-evaporitic assemblages, generally. This first index is adopted as most workable, thereby placing 400 feet of beds in the Halfway Formation. Minor anhydritic beds extend a further 100 to 150 feet below this boundary.

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* The specific name of this fossil is accepted from Böhm (1903) without revising the complicated synonymy.

Exposures of the Halfway occur about the footwall of the Solitude thrust and along the Hart Highway 0.7 mile northeast of Silver Sands Creek. In the latter, dark-coloured clastic-carbonate strata are shown in the core of an unnamed anticline. The beds here yielded *Lingula* sp. and pelecypods sp. indet.

Charlie Lake Formation.—Little of the Charlie Lake Formation can be appreciated from exposure, mostly as a result of its deep weathering. A complete section of about 475 feet, stratigraphic interval, is found in drill cuttings from Triad Bush Mountain b-23-A (1), and this may include some repetition by faulting, probably less than 40 feet. Drill cuttings from Triad B.P. Bush Mountain a-15-A cover the upper 400 feet of the formation.

The Charlie Lake Formation is an assemblage of carbonates, clastics, and anhydrites. It is notable for a wide range of lithology and the prevalence of buff, white, and light colours. The carbonates include aphanitic and crystalline limestones and dolomites; cryptocrystalline dolomites, which may be of primary deposition; limestones, and dolomites with drusy, and sub-oolitic textures; soft white limestones with anhydritic intergrowths in places; arenaceous limestones and dolomites; grey limestones of Baldonnel types, which are included in the upper part; and dark-grey dolomites and anhydritic dolomites, which mark a transition to the Halfway, in the lower parts. The clastics are sandstones with calcareous, dolomitic, quartzitic, and anhydritic matrices; quartzites; calcareous siltstones, in part anhydritic; buff and light-grey shales, seen in few thin interbeds. Sandstones are very fine to medium grained, clean and without argillaceous content. Clastics and carbonates make up most of the formation in drill cuttings. Anhydrite forms less than one-tenth of the formation. This fraction also includes some gypsum and gypsiferous material, thought to be the result of hydration and alteration of anhydrite by weathering, drilling, and the treatment of drill cuttings. Anhydrite of white crystalline and platy appearance, and as intergrowths and vein material, occurs in drill cuttings. Anhydrites are common in the lower 125 feet of the formation. Details of the occurrences are perhaps confused by minor faulting in Triad Bush Mountain b-23-A (1).

A threefold division of the Charlie Lake Formation is apparent in the subsurface sections.

(*c*) Upper Beds: Grey limestones; light-coloured limestones and dolomites; lesser clastics; few anhydrites (mostly in the lower 150 feet); thickness, about 260 feet.

(*b*) Middle Beds: Sandstones for the most part, with thin silty, anhydritic, and calcareous layers; thickness, about 90 feet.

(*a*) Lower Beds: Light-coloured limestones and dolomites; clastics; anhydrites; with dark-grey dolomites and anhydritic dolomites in lower half; thickness, about 125 feet.

The lower beds (*a*) are the most anhydritic.

A transition relates the Charlie Lake to the overlying Baldonnel Formation. The boundary is drawn to restrict light-coloured carbonates to the Charlie Lake Formation. Association of these carbonates with clastics and anhydrites, lower in the sequence, supports the assignment.

Road cuts along the Hart Highway, 0.2 to 0.5 mile northeast of Silver Sands Creek, reveal parts of the Charlie Lake Formation. Sandstones of the middle beds (*b*), in exposure of 60 to 70 feet, stratigraphic interval, in the west limb of the

Silver Sands anticline, provide a datum for the field divisions of the Grey Beds. Away from the road cuts, the boundary of the field divisions (i) and (ii) is indicated by light-coloured limestones and dolomites seen in float.

The Charlie Lake beds have not yielded fossils locally.

The Baldonnel Formation.—Most of the Baldonnel Formation consists of limestones; dolomitization is not extensive or advanced in the area of the western outcrops, and clastics are of subordinate order. Only lower parts of the formation were cut in drilled sections. Drill cuttings therefrom include grey limestones, of argillaceous, silty, and shell fragmental types, and few recrystallized and partially dolomitized limestones. Outcrops provide the most complete record of the Baldonnel Formation, as given for the uppermost 500 to 550 feet of the Grey Beds, the upper field division (ii). Exposures of the Baldonnel are small. Those near the Pacific Great Eastern Railway bridge at Mountain Creek allow the best view of the formation.

Benthonic shelly faunas are characteristic of the Baldonnel Formation. Of its fossils listed in the field description, upper field division (ii), *Lima? poyana* from the lowermost 100 feet and *Dielasma? (T.)* cf. *julica* from the middle of the formation are most significant in regional correlations.

Equivalence of field divisions of the Grey Beds and formations observed in drilled sections can be summarized: Division (i), the lower field division, is equivalent to the Halfway and the lower and middle beds of the Charlie Lake Formation; Division (ii), the upper field division, is equivalent to the upper beds of the Charlie Lake and the Baldonnel Formation.

Stratigraphic work by McLearn established a faunal sequence of the Grey Beds: in the lower part, the upper range of the *Nathorstites* fauna; the Mahaffy Cliffs, and the Red Rock Spur fauna near the middle; and the *Lima? poyana* fauna near the top. The Grey Beds belong partly to the Karnian stage. Their lower part, which together with the underlying Dark Siltstones bear the *Nathorstites* fauna, are probably Ladinian (Middle Triassic) in age (*see* McLearn, 1947, and work cited by Tozer, 1961, p. 9). The Anisian *Beyrichites-Gymnotoceras* fauna of the Toad Formation sets a lower limit to the dating of the *Nathorstites* fauna. For later notes on the stratigraphic placing of these faunas in northeastern British Columbia, *see* Hunt and Ratcliffe (1959), Tozer (1961), Armitage (1962), Colquhoun (1962), and Pelletier (1964).

The *Nathorstites* fauna and the Mahaffy Cliffs and Red Rock Spur fauna have not been found in the Pine Valley. The shelly faunas of the Baldonnel Formation, in the upper third of the Grey Beds, in Division (ii), may be ascribed to the Karnian stage of the Upper Triassic according to occurrence of *Lima? poyana* McLearn, the knowledge of its stratigraphic relations to ammonite faunas in the Foothills area of the Peace River (McLearn, 1960), and the occurrence of *Dielasma? (T.)* cf. *julica* (Bittner). It is noted that the morphology of the latter species and its range of variations may not be fully proscribed. *Pteria? (A.) bittneri* was first described from Bear Island (Böhm, 1903). Its age, and that of its associated fauna or its separate elements, remains in doubt—Ladinian or Karnian or both. The associated fauna was reported to contain *Nathorstites mcconnelli* var. *lenticularis* Whiteaves and *Dawsonites canadensis* Whiteaves (the two ammonites are representatives of the *Nathorstites* fauna of northeastern British Columbia) together with *Trachyceras* and *Halobia*. The *Myophoria*'s with distinct carina, pro-carinate sulcus, and cancellated ornament of the ribbed pro-sulcate sector, as in the species noted from the Pine Valley, and also *Septocardia's* are common in benthonic faunas of later Middle and Upper Triassic age.

The Halfway and Charlie Lake Formations can only be assigned a Ladinian-Karnian age, the dating Middle or Upper Triassic being left unresolved in the map-area.

The Ladinian-Karnian sedimentary environments consisted of a cratonic shelf in the area of the Plains, bordered by a miogeosynclinal slope (or shelf) on the west, along the site of the Rocky Mountains. Carbonates, clastics, and evaporites accumulated on the cratonic shelf. Evaporitic and anhydritic sequences are relatively thick under the Plains. Sands and silts reworked and transported across the cratonic shelf contributed to the sediments of the Grey Beds laid down on the miogeosynclinal slope. The prevailing sedimentary transport was from the northeast (Pelletier, 1964). In the Foothills, the Charlie Lake Formation represents a westward extension of the cratonic shelf environment. The anhydrites were deposited in hypersaline waters of restricted circulation, probably brought about by local sand accumulations and intermittent tectonic warping. Light colours of the Charlie Lake beds suggest the abeyance of reducing conditions on the sea bottom—unlike conditions for most of the Triassic of the Foothills. A return to normal marine conditions represented by the Baldonnel limestones preceded an eastward shift of the miogeosynclinal slope. Finer clastic components, and the dominance of pelagic faunas in the Pardonet, record this later change.

Pardonet Formation. — McLearn (1940) first recognized the Pardonet, the uppermost lithological unit of the Triassic in the Peace River Foothills.

The formation is present in the Pine Valley. Here, it overlies the Grey Beds conformably, and the contact is a gradation. The boundary is drawn where siltstones, sandstones, arenaceous and shell fragmental limestones give way to argillaceous limestones and limestones with *Halobia*. The Pardonet consists of limestones which are argillaceous or silty to some varying degrees. Aphanitic limestones occur with *Halobia* beds; argillaceous limestones become more common upwards; and the uppermost beds with *Monotis* are notably silty. Some recrystallization with partial dolomitization in few beds appears to be local and of minor order. The Pardonet limestones are dark grey and have a fetid sulphurous odour when broken. They weather in slabs with shaly and silty aspect, and in grey to purplish-brown colours.

The Pardonet is very fossiliferous. Pelecypods were most abundant. Species of *Halobia* and *Monotis* formed large populations. Their remains now make beds of layered, densely packed and compressed shells. The extent and nature of the beds and the thin shells indicate the former pelagic habit of *Halobia* and *Monotis* (*see also* Tozer, 1961). Intervals between the shell beds have dispersed fossils. *Halobia* occurs in the lower part of the Pardonet; *Halobia* with *Oxytoma* cf. *O. mucronata* Gabb in the middle, and in the upper part *Monotis* including a shell band of *Monotis subcircularis* Gabb. Several species of *Halobia* are present, and include the *cordillerana* Smith, *dilatata* Kittl, *pacalis* McLearn, *lineata* Münster, and *ornatissima* Smith forms or affinities. Swarms of fossil immature *Halobia* occur at some horizons. Westermann (1962) zoned *Monotis* beds for one local section at the West Pine Bridge by recognizing eight species and four subspecies of *Monotis*. The species *M. subcircularis* Gabb seems the most significant in field use. Parts of the generalized *Monotis* zone are exposed frequently or shown by float—a distinct advantage in mapping. Other pelecypods are less abundant in the Pardonet: *Pleuromya*? *madisonensis* Smith found in the lower part about the transition to the Grey Beds; *Gryphaea* sp., possibly *G. chakii* of McLearn from the middle; *Lima* sp.; *Entolium* sp.; and pectinids not identified. Ammonites seem rare in contrast to their abundance in parts of the Peace River Foothills, though scarcities and different abundances of ammonites from place to place can be noted from McLearn's observations there (1960). Ammonites from the Pine Valley are *Juvavites kellyi* Smith, *Juvavites magnus* (*-concretus*) McLearn, *Juvavites* sp., *Malayites* sp., from the interval of *Halobia–Halobia* with *Oxytoma* cf. *O. mucronata*. The nautiloid cf. *Proclydonautilus natosini* McLearn came from this interval, and *Clydonautilus* sp. from the lower part of the *Halobia* interval and about 250 feet above the base assigned to the Pardonet on the east side of Silver Sands Creek. Reptilian remains are known in the Pardonet, and black fossilized pieces of bone are often found.

The Pardonet Formation is 700 feet thick, from measurements on the east side of Silver Sands Creek. The sequence of pelecypod faunas can be traced out here, from small exposures and remanié. The *Halobia* beds are exposed at Cairns Creek about 1 mile upstream from the highway. Railway and road cuts at the West Pine Bridge show the *Monotis* beds folded in a tight anticline of concentric form. In the west limb of the anticline, a concealed interval of 6 feet separates *Monotis* beds from overlying strata of the Fernie Group. There is a second anticline, of angular form, directly overthrust on the anticline exposed in the railway and road cuts. Pardonet beds of the *Halobia–Halobia* with *Oxytoma* cf. *O. mucronata* sequences occupy the core of the second, upper anticline. In its west limb, *Monotis* beds in vertical attitudes adjoin the Fernie outcrop of the underthrust in the valley of Silver Sands Creek. The thrust fault separating the anticlines dips 30 to 40 degrees southwest. A fault (possibly the thrust fault noted) cuts off Fernie beds in the west limb of the anticline exposed in the railway and road cuts. Pardonet beds with steep to nearly vertical dips come against the fault on the west. Fault planes are not directly visible, and the relation of the faults are not clear on the ground.

The Pardonet beds are Upper Triassic in age. According to McLearn's study of the ammonite faunas and their correlations (1960), they include parts of the Karnian and Norian Stages. The pelecypods support this dating. Species of *Halobia* and *Monotis* have a world-wide distribution, and these pelecypods attained their acme in the Upper Triassic. *Monotis* faunas were widespread in the Norian of the Pacific, Himalayan, Alpine, and Arctic provinces. The ammonite *Juvavites kellyi* Smith belongs to the uppermost Karnian, and the *Juvavites* subzone (later renamed the *Tropites welleri* subzone by Silberling, 1959) in the highest range of the *Tropites subbullatus* zone. Much of the combined *Halobia–Halobia* with *Oxytoma* cf. *O. mucronata* interval falls in the Karnian Stage. *Juvavites magnus* McLearn and its variant *concretus* are reported from the *Pterocera–Cyrtopleurites magnificus* beds in the Peace River Foothills, and may be assigned either to the lowermost Norian following McLearn (1960) or to the uppermost Karnian.

It is uncertain whether the uppermost part of the Norian Stage is represented in the Pardonet Formation. The Rhaetic Stage is apparently missing, and is unknown for the Rocky Mountain Foothills in general. Jurassic strata, the Nordegg limestones, follow closely on the Monotis beds, as noted at the West Pine Bridge. Nordegg limestones lie 40 to 60 feet, stratigraphic interval, above beds with *Monotis subcircularis* at Silver Sands Creek and west of Cairns Creek.

JURASSIC

The Jurassic System contains strata of the Fernie Group and the lower part of the Beaudette Group (the Lower Marine Bullhead).

Fernie Group

The term Fernie is now widely accepted in northeastern British Columbia, following the recommendation of Hage (1944). Strata of this name extend northward from the central and southern Rocky Mountain Foothills and are present in the subsurface of the plains without much change in lithology. Another term, Pine River, is a discarded synonym, and was first applied to Cretaceous shales in the Pine Valley (Spieker, 1921). Group status accorded to the Fernie by Frebold (1957) leads to systematizing its lithological divisions as formations, but this nomenclature remains unsettled. Accordingly, provisional stratigraphic terms are used here. Three divisions can be made for the Fernie beds in the Pine Valley, though others may be resolved by detailed studies. The lowest is the Nordegg, represented here by limestones and clastics with chert. The overlying shales, with few silty and sandy parts, are described by the term Middle Shales. The third, upper division is the Transition Beds—interbedded shales, siltstones, and sandstones; these mark the change to the sand deposition of Beaudette times.

The Fernie Group is 650 to 800 feet thick. Its largest stratigraphic division is the Middle Shales, between 500 and 600 feet thick. The lower and upper divisions have narrow outcrops, and are not distinguished by topographic relief. Their distribution in relation to the whole of the Fernie Group makes it unnecessary to represent them separately in the mapping.

The lowest Fernie beds overlie the Pardonet in parallel order, but the contact is concealed. The Pardonet Formation thins eastward under the Foothills, and is absent under the Plains where erosion preceded Jurassic deposition. Fernie strata reflect comparatively stable tectonic conditions for much of the Jurassic. Frebold's work (1957) showed evidence for non-sequences, thin reduced sedimentation and intervals of non-deposition in the Foothills, southern Alberta. Zonations of this order are not available for the Fernie Group in northeastern British Columbia.

Nordegg Beds.—The lowest beds seen in the Fernie Group are black argillaceous and fine crystalline limestones with thin interbeds and partings and layers of calcareous shale. The limestones contain *Oxytoma* sp., *Chlamys* sp., *Gryphaea* sp., and a rhynchonellid brachiopod. Later work, the identifications of *Oxytoma* cf. *O. inequivalvis* (Sowerby), *Chlamys* n. sp. aff. *C. textorius* (Münster), *Entolium* cf. *E. calvum* (Goldfuss), and *Furcirhynchia striata* Quenstedt, by Ager and Westermann (1963), supports the reference to the Nordegg and indicates a Lower Jurassic (probably Sinemurian) age. The limestones are overlain by thin interbedded dark-grey shales, siltstones, and sandstones, with limestone layers and thin chert banding.

The Nordegg is 50 to 60 feet thick at the West Pine Bridge. Here, the lower contact with the *Monotis* beds lies in a 6-foot concealed interval, the lower limestone member is 33 feet thick, and the upper clastic member 17 feet thick; dark-grey shales define the top of the formation. The exposure is in the west limb of the anticline along the line of the Pacific Great Eastern Railway (*see under* Pardonet Formation). Samples of limestones and shales from the Nordegg of the exposure have slight to low proportions of phosphate. Exposures of the Nordegg limestones are also known from the Silver Sands and Cairns Creeks.

Middle Shales.—The Middle Shales contain black and dark-grey shales, fissile and concretionary shales, shales with organic markings (small "worm tracks"), and dark-grey mudstones, argillaceous sandstones in minor amounts, and some clay ironstone banding. Fossils are not plentiful, and those noted in the field were *Cucullaea* sp., *Gryphaea* sp. (fragments), *Pleuromya?* sp., pectinids, and belemnite fragments. The shales are soft, easily weathered, and rotted. They were incompetent in the deformation, and many faults cut their outcrops. Most exposures show some structural disturbance, for example those on the north salient of Mount Le Moray.

The Middle Shales are about 525 feet thick by estimate, or not more than 600 feet by mapping. They can be dated Lower to Upper Jurassic by stratigraphic position.

Transition Beds.—In the upper part of the Fernie, interbedded dark-grey shales, silty shales, siltstones, and sandstones form a transition to the overlying Beaudette Group. They are sufficiently distinct to describe as a stratigraphic unit, the Transition Beds. In exposures along the Pyramis thrust north of the Pine River, 75 feet of beds are assigned to this unit. Exposures elsewhere are incomplete, and estimates allow a thickness to 150 feet farther west in the map-area, though the outcrops may be disturbed by faulting. The proportion of sandstones increases upwards in the Transition Beds. Sandstones are fine grained and argillaceous. They form beds 1 to 40 inches thick, that weather in rusty-brown colours. The sandstones yield small forms of *Buchia* sp. and small pentacrinoid ossicles with decorated disks. The Transition Beds appear to vary, from place to place, in pattern of interbedding and in the proportion and bed thickness of sandstones. Such differences may reflect variable deposition or perhaps some stratigraphic discordance. The observed contacts of the Transition Beds are gradational, and are designated arbitrarily, the lower contact by the lowest sandstone layers, and the upper contact with the Beaudette Group at the base of sandstone beds, 10 feet or more thick. Shales and shales with siltstones and thin sandstones continue into overlying parts of the Beaudette Group.

A possible dating for the Transition Beds is obtainable from fossils identified by Jeletzky (1962). They included:—

(i) *Buchia mosquensis* (Buch, non Keyserling, non Lahusen) (late form).
Buchia piochii (Gabb)?
(ii) *Buchia* ex gr. *B. mosquensis* (Buch)–*B. concentrica* (Sowerby).
(iii) *Pecten* (*Entolium*?) sp. indet.
Pteria (*Oxytoma*) sp. indet.

The three lots were from float of local or exotic origin in Little Boulder Creek. The fossils are referred to the Transition Beds which are exposed in places along the creek wall below thick sandstones of the Beaudette Group. Jeletzky (1962) suggests a Portlandian age for the *Buchia mosquensis* (Buch) in lot (i), because it is a late form of the species. The *Buchia mosquensis* zone includes the late Kimeridgian and most of the Portlandian Stage (*sensu stricto*), excepting the uppermost Portlandian. The lots (i) and (ii) may be of the same age. However, an early Kimeridgian or Upper Oxfordian dating for the fossils of lot (ii) cannot be excluded, as they are poorly preserved.

The Transition Beds are assigned to the Upper Jurassic, to the Oxfordian-Kimeridgian-Portlandian or some parts of this interval. Overlying sandstones in the lower part of the Beaudette Group, and in the Monteith Formation, are also assigned to the Upper Jurassic. Faunas corresponding to the latest Upper Jurassic, late Upper Tithonian in the terms applied to the Western Tethyan province, also the late Portlandian and Purbeckian of the standard Northwest European succession (Arkell, 1956), are seemingly absent from the Transition Beds. The Transition Beds can be correlated with the Passage Beds of the Alberta Foothills, or with part of this sequence. Also, Jeletzky (1962) proposed the correlation of beds with *Buchia mosquensis* (Buch, non Keyserling, non Lahusen) and *Buchia piochii* (Gabb)? in the lot (i) above, with the lower part of the Nikanassin Formation.

JURASSIC AND CRETACEOUS

Strata in the interval between the Fernie and Fort St. John Groups have been described under the name Bullhead. They comprise two natural divisions of major order: in the lower part—marine beds, sandstones, quartzites, siltstones, and shales; in the upper part—coal-measure sequences. The divisions were named the Beaudette and Crassier Groups respectively (Hughes, 1964). The classification of the Bullhead strata (ibid.) carried proposals that the term Bullhead be kept to its original meaning and definition (*see* McLearn and Kindle, 1950, p. 63), denoting a supergroup or succession, and that the Dunlevy Formation of Beach and Spivak (1944), which includes all formations of the Beaudette Group and the lower two formations of the Crassier Group, be discarded. The Beaudette and Crassier Groups outcrop in the Inner Foothills of the map-area. They underlie the surface of the Outer Foothills and of the Plains.

Beaudette Group

The Beaudette Group contains 600 to 3,500 feet of strata, in the Foothills area of the Peace, Moberly, and Pine Rivers. The thickness increases from northeast to southwest. In the eastern and middle parts of the Foothills the Beaudette strata form a continuous sequence of deposits which can be divided into three formations—the Monteith, Beattie Peaks, and Monach, in ascending order. To the southwest, at least locally in the Pine Valley, because of facies changes they are unobtainable or become uncertain, and so the Beaudette strata are best mapped as an undivided unit. The Monteith Formation consists of sandstones and quartzites for the most part; the Beattie Peaks Formation, of interbedded shales, siltstones, and sandstones; and the Monach Formation, of sandstones with or without quartzites which have variable distribution in its upper part. From northeast to southwest, across the Foothills, sandstones replace quartzites in the Monteith Formation, the proportion of sandstones increases in the Beattie Peaks Formation, and the Monach Formation thickens and persists with little difference in lithology. Such change of shale-siltstone-sandstone-quartzite ratio is interpreted from the available exposures. The Monteith, Beattie Peaks, and Monach Formations have their typical development in the middle part of the Foothills, more especially in the Carbon Creek-Moberly River area, where Mathews (1947) first named them.

The thick Beaudette strata were laid down in a subsiding trough. They were partly the result of Nevadan tectonism, uplift and erosion to the west, and a redevelopment of the miogeosyncline about the site of the Rocky Mountain Foothills. Much of the Beaudette sediments came from westerly sources, according to the evidence of feldspar and chert noted in the lower part of the group, the distribution and thickness of the sandstones, and a common east-northeast dip and set of current bedding. However, the distribution of the Monteith quartzites suggests that they were partly derived from easterly sources.

Undivided Beaudette Group.—Beaudette strata are mapped as an undivided group about the Pyramis thrust and farther southwest. Sandstones in thick beds make up most of this unit. The base of the lowest sandstone, 10 feet or more thick, serves to mark the Fernie-Beaudette boundary. Post-Cretaceous erosion has removed part of the undivided Beaudette Group in most outcrops. The west limb of the Pyramis syncline has the most complete sequence, 3,150 feet of beds, mostly sandstones as seen in exposures north of the Pine River. The upper contact is faulted here, or is disturbed by small scale folding.

The undivided Beaudette Group has grey and brown, fine- to medium-grained argillaceous sandstones of feldspathic and ferruginous types in its lower and middle parts, and cleaner, lighter-coloured sandstones with some quartzitic sandstones in the middle and upper parts. Quartzites occur in one bed 40 feet thick, about 1,650 feet above the base of the group in the Pyramis syncline north of the Pine River. Quartzitic and argillaceous sandstones are present about this stratigraphic level, 1,500 to 2,000 feet above the base of the undivided Beaudette Group, in the outcrop area. Beaudette sandstones form beds 5 to 20 feet thick. They are separated by black or brown shales with partings and layers of siltstone and sandstone, in units 1 to 30 inches thick. Shales, siltstones, and grey quartzitic sandstones containing particles and flakes of black shale form minor components of the lithology. The sandstones have current bedding of long sweep, and rest on distinct bottom sets or on shales. The latter contacts are marked by load casts, flow casts, impress and flame structures.

In places, Beaudette sandstones contain pebbly layers and layers of shell fragments. Layers of clay galls and of shale and mudstone inclusions are features of the sandstones. These are common in the interval 1,800 to 2,250 feet above the base of the Beaudette Group at the north salient of Mount Le Moray. Clay galls have simple oval outlines with median diameter one-quarter to 1 inch, and flattened, discoidal, lenticular shapes. They are dispersed and separated in the sandstones. Shale and mudstone (or argillaceous) inclusions have diffuse, irregular, or angular outlines which may grade or penetrate into the sandstone matrix, and are about the same size as the clay galls but thicker. The shale and mudstone inclusions are dispersed, or more concentrated, and about contiguous in some layers.*

The north salient of Mount Le Moray provides a sequence of the lower 2,250 feet of the undivided Beaudette Group. Here, a lower fossiliferous interval of 1,525 feet contains brown-weathering, argillaceous, feldspathic, ferruginous sandstones in thick beds; concretionary banding, layers of clay galls, shell fragments, and, more rarely, pebbly layers are features of the sandstones; shales in thin interbeds form a minor proportion of the exposures. This lower interval yielded pelecypod faunas in the following collections: Lot (i), from float about 25 to 100 feet above the base of the Beaudette Group—*Buchia* sp. indet. (of uppermost Jurassic or lowermost Cretaceous affinities), *Tellina*? sp. indet.; lot (ii), float about 1,000 to 1,075 feet above the base—*Buchia* ex aff. *B. fishcheriana* (d'Orbigny) and *B. okensis* (Pavlow), *Buchia* n. sp. ex aff. *B. volgensis* (Lahusen); lot (iii), float about 1,300 to 1,350 feet above the base—*Buchia* cf. *B. okensis* (Pavlow), *Buchia* cf. *B. uncitoides* (Pavlow) s.l., *Buchia* sp. indet., *Arctica*? sp. indet.; lot (iv), float 1,350 to 1,500 feet above the base—*Buchia* sp. indet. (of early Lower Cretaceous affinities), *Buchia*? sp. indet., *Pecten* (*Pseudamusium*?) sp. indet. Sandstones continue in the overlying interval of 275 feet, to 1,800 feet above the base of the Beaudette Group: these are buff- and brown-weathering sandstones and light-grey quartzitic sandstones; the sandstones are clean and less argillaceous than for underlying intervals; some argillaceous sandstones with few layers of clay galls, and shale and mudstone inclusions are present in the upper part. In the uppermost 450 feet of section the lithology is complex, and in the available exposures includes brown-weathering, fine- to medium-grained argillaceous sandstones with chert grains; quartzitic sandstones, variably argillaceous; quartzitic grits; brown and dark-grey shales; siltstones and silty mudstones in thin beds and layers, associated with shales. Many of the sandstones contain angular to rounded inclusions of shales and mudstones, close packed and in close bedded layers, and have a conglomeratic appearance. Layers of dispersed clay galls are also present. The remainder of the sequence is missing due to post-Cretaceous erosion. No fossils were obtained above the collection, lot (iv).

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* Field observations lead to the views:—

Clay galls were developed by break-up, scour, and dispersion of thin mud layers preceding deposition of sands; the clay galls probably remained in place or were shifted to and fro. Argillaceous inclusions, in continuous layers or in lenses, were formed by plastic flow and break-up of mud layers under sedimentary loading, and by incorporation of the phenoplasts in the yielding sands. A similar origin can explain the argillaceous inclusions of disperse habit. These inclusions were separated by scour and erosion, with or without transport. In some examples, plastic deformation under load modified their shapes. (Lenticular, argillaceous inclusions of disperse habit include few which have sandy and silty laminæ; the laminæ are deformed, or inclined to the bedding of the formation.)

Clay galls occur throughout the marine and non-marine sandstones of the Cretaceous. The argillaceous inclusions, phenoplasts or phenoclasts, are less common: those in concentrated or contiguous habit are known from the uppermost beds of the Beaudette Group in the incomplete section at the north salient of Mount Le Moray; those in disperse habit occur in the Beaudette Group and in the post-Beaudette strata, where they seem restricted to non-marine sandstones.

Jeletzky (1962) identified the faunas, and his observations concerning their dating are summarized here. The lower faunas, (i) and (ii), can be dated as Upper Jurassic or Lower Cretaceous (late Tithonian and early Berriasian Stages respectively). The third fauna, (iii), belongs to the basal Lower Cretaceous (early Berriasian).

A fauna containing *Buchia* cf. *B. uncitoides* (Pavlow) and *B.* ex aff. *B. keyserlingi* (Lahusen) (small early forms) was obtained about 1,500 feet above the base of the undivided Beaudette Group, in the west limb of the Pyramis syncline, north of the Pine River. Jeletzky (1961) suggested a Lower Cretaceous (late Berriasian) age for these fossils; their preservation does not allow a definite dating.

An Upper Jurassic and Lower Cretaceous age is assigned to the undivided Beaudette Group. The Jurassic-Cretaceous boundary may be placed about the level of the fauna with *Buchia* ex aff. *B. fischeriana* (d'Orbigny) and *B. okensis* (Pavlow), and *Buchia* n. sp. ex aff. *B. volgensis* (Lahusen), or in the lower 1,000 feet of the undivided Beaudette Group. The lower 1,500 to 2,000 feet of the undivided Beaudette Group is correlative with the Monteith Formation of the eastern outcrops, according to lithology, stratigraphic position, and the limits given by the faunal dating. The upper parts of the undivided Beaudette Group are then equivalents of the Beattie Peaks and Monach Formations.

Monteith Formation.—The Monteith Formation consists of sandstones, quartzitic sandstones, and quartzites, together with minor thin beds of shales and shales with siltstone and sandstone partings. The thickness of the formation is about 1,000 to 1,850 feet, increasing from east to west across the Foothills of the Peace and Pine River areas. The Monteith Formation includes 1,500 to 1,550 feet of beds in the Pine Valley, measured across the west limb of the Bickford anticline. Shales in a small exposure in the core of the anticline north of the Pine River may belong to the Fernie Group. The lower boundary of the Monteith is unseen in the map-area, but to the north the formation is reported to overlie the Transition Beds of the Fernie Group.

The Monteith sandstones are grey and brown in colour, fine to medium grained, and argillaceous, feldspathic, and ferruginous to varying degrees. They resemble those of the undivided Beaudette Group in lithology, bedding, and sedimentary textures. The quartzites are fine to very coarse grained, and occasionally have thin layers of quartzitic grit and pebbles. The sandstones and quartzites are interbedded. They represent two lithofacies which interfinger and replace each other in the Monteith Formation. Quartzites and quartzitic grits and sandstones are abundant and distributed throughout the formation in the eastern Foothills of the Peace River area; they are restricted to its upper third part in the middle ranges of the Foothills. From Falls to Big Boulder Creeks in the Pine Valley, quartzites with sandstones occur in the upper 300 to 500 feet of the Monteith Formation, but their development seems irregular, and their proportion variable. Farther west in the Pine Valley, quartzites form a very small part of the undivided Beaudette Group.

The Monteith Formation can be dated as Upper Jurassic and Lower Cretaceous age (late Tithonian, Berriasian, and possibly early Valanginian) from its stratigraphic position. It has yielded a few fossils, among which *Buchia* sp. was identified by Jeletzky. Comminuted plant debris occurs in the sandstones and quartzites, as commonly in other Cretaceous marine beds.

CRETACEOUS

Cretaceous strata form most of the outcrop in the Foothills area of the Pine Valley. They include the major part of the Beaudette Group, the Crassier and Fort St. John Groups, and the Dunvegan Formation.

Beattie Peaks Formation.—The Beattie Peaks Formation consists of thinly interbedded black and dark-grey shales, silty shales, siltstones, thin sandstones, with ironstone banding, together with sandstones in beds 4 to 20 feet thick. Sandstones are fine to medium grained, buff and brownish coloured; in places they have layers of shell fragments, layers of *Dentalium*? sp., and layers of clay galls. Organic (or worm) casts and markings are common in the thin-bedded shales and siltstones.

The Beattie Peaks overlies the Monteith Formation conformably, and the contact is abrupt, or marked by a brief change of lithology. The thickness of the Beattie Peaks Formation is 650 feet in the east limb of the Bickford anticline, and increases westward to perhaps 950 feet at Big Boulder Creek, where its upper boundary is not well defined. The proportion of sandstones in thicker beds increases in the upper half of the formation, partly making a transition to the overlying Monach Formation. Outcrops of the Beattie Peaks Formation are indistinct and not clearly separable from those of the Monach, where exposures are slight; sandstones are prominent, and they are similar in both formations.

Advice and identifications by Jeletzky (1957, 1961, 1962) indicate a Lower Cretaceous (Valanginian) age for the Beattie Peaks Formation. *Dentalium*? sp. seems a useful, local guide fossil for identifying the formation. Only the *Buchia*'s and ammonites are of general stratigraphic value. They are listed: *Buchia keyserlingi* (Lahusen) f. typ., *Buchia keyserlingi* (Lahusen) var. (*B. crassicollis*-like variant), *Buchia* cf. *B. keyserlingi* (Lahusen), *Buchia sublævis* (Keyserling), *Buchia* cf. *B. terebratuloides* Lahusen, *Polyptychites* (*Polyptychites*) cf. *P. keyserlingi* Neumayr and Uhlig 1881 s.l., *Dichotomites* aff. *D. quatsinoensis* (Whiteaves) and *D. oregonensis* Anderson.

Monach Formation.—In the first account by Mathews (1947), the term Monach referred to sandstones. It was then extended to overlying and associated quartzites (Hughes, 1964). The Monach Formation is between 300 and 400 feet thick in the Bickford anticline, and perhaps less farther west, where exposures do not allow a clear distinction from the upper part of the Beattie Peaks Formation. The Monach Formation is divided into two members:—

Member (i) (The Lower Sandstones): This division consists of argillaceous and cleaner quartzitic sandstones, mostly medium grained and current bedded. Shales with siltstone and sandstone, in intervals less than 6 feet thick, are a minor component, less than one-tenth of the formation. The sandstones weather in light buff and brownish colours, often with a flaggy or platy habit. They form thick beds, and in places contain lenses of quartzitic grit, and fragments of shells and belemnite guards, and clay galls in layers. The fauna *Buchia* n. sp. aff. *B. inflata* (Toula), and *Dichotomites* s.l. cf. *D. giganteus* (Imlay) from the Lower Sandstones, Member (i), was identified by Jeletzky (1961) and dated Lower Cretaceous (mid to late Valanginian). Member (i) is 335 feet thick where exposed in the east limb of the Bickford anticline, at the Pacific Great Eastern Railway cut 2 miles southwest of Beaudette Creek.

Member (ii) (The Upper Quartzites): The quartzites have variable development. They are thin (2 to 3 feet) or absent in some places in the west limb of the Bickford anticline; elsewhere they form a single unit 20 to 45 feet thick; or they may occur as several beds with concealed parts in intervals from 50 to 130 feet thick, as in the ground between Fisher and Crassier Creeks and on the west limb of the Big Boulder Creek anticline south of the Pine River.

In the railway cut 2 miles southwest of Beaudette Creek, 42 feet of quartzites rest on a sharp, plane contact with the sandstones, and are overlain directly by coal measures of the Brenot Formation. Elsewhere the field relationships are partly concealed and obscure.

The Monach quartzites are fine to very coarse grained, with occasional lenses of quartzitic grit. Weathered surfaces are often porous, and cementation is incomplete in patches. By their resistance to erosion and frequent exposure, the quartzites form a useful though inconsistent marker for mapping the Beaudette-Crassier boundary, but contacts are usually concealed. The quartzites are unfossiliferous.

The Monach Formation represents filling of the geosynclinal trough as the seas retreated in later Valanginian time. The quartzites of Member (ii) are regarded as littoral and near-shore sediments abandoned by the retreat of the sea. Possibly they were reworked before the onset of Crassier sedimentation, and in places interlayered with marine or non-marine beds before coal-measure sedimentation was established and continuous in Crassier time. In other views, the quartzites of Member (ii) may be assigned to the Crassier Group or treated as an independent stratigraphic unit.

Crassier Group

The Crassier Group are coal measures containing mudstones, shales, siltstones, sandstones, sandstones with grits and conglomerates, and coals. The group is divided into the Brenot, Dresser, and Gething Formations in ascending order, on the basis of their shale/sandstone ratios, differences in the cyclothems, relative thickness of coal seams, and occurrences of medium- to very coarse-grained sandstones, grits, and conglomerates. Three divisions were first established in the Fisher Creek area: lean, partly barren, and thin-bedded coal measures in the Brenot Formation; coal measures with much medium- to very coarse-grained sandstones, and lesser grits and conglomerates in the Dresser Formation; coal measures with a lesser proportion of sandstones, and well-developed cyclothems and coal seams in the Gething Formation. The same sequence continues northeastward to the Foothills area of the Peace River, where the Dresser Formation corresponds to the upper 700 to 900 feet of the Dunlevy Formation of Beach and Spivak (1944), and McLearn and Irish (1944).

The Crassier Group is between 3,500 and 3,750 feet thick in the Pine Valley. It is disturbed by faulting and folding in many places. The outcrop has little distinctive topography, although the Dresser Formation shows some relief as its sandstone beds form ridges on valley slopes and high ground. Much of the Crassier outcrop is forest covered and parts are obscured by drift. Formational divisions are not easily obtained where structures are complex and details concealed. The main difficulty is placing the gradational boundaries everywhere reliably. The Brenot, Dresser, and Gething Formations are separable east of the Bickford anticline. However, in several places complex structures and lack of exposures make it more practicable to show the coal measures of the Crassier Group as an entire unit. One complete unfaulted section of the formations in the west limb of the Fisher anticline makes a useful reference: parts are visible along tributaries entering Fisher Creek from the east, 2 to 4 miles north of the Hart Highway (Hughes, 1964; Figs. 2 and 8). Formations of the Crassier Group are not well differentiated farther west, where there is a general increase of sandstones. The coal measures are better treated as an undivided group west of the Bickford anticline.

The Crassier coal measures record past environments of shallow lakes, swamps, and forests. The sediments filled a continually subsiding trough. They were deposited in repeated cycles and now consist of mudstones, shales, sandstones, lithified seat earths and underclays, coals, and roof shales, or in partial cycles of this order. Mudstones, shales, and sandstones represent the former lacustrine sediments; lithified seat earths and underclays represent the sediments of swamps and their modification by organic activity, or partly or less certainly, fossil soils; coals represent the swamp and forest growth until subsidence and flooding at the start of the next cycle; and sandstones, grits, and conglomerates, the deposits of rivers and deltas crossing over the site of the Foothills. Seemingly, most strata of the cyclothems are non-marine. A few marine phases are recorded or indicated; for example, a cyclothem of the Gething, and a thin quartzite in the Dresser Formation (depth 3,630 to 3,640 feet, in the French Petroleum Company-Richfield Brenot Creek No. 1 well in the Peace River area). Crassier sedimentation shows much local variation—differences in the thickness of sandstones and coals, their limitations in area, splitting of coal seams, washouts, and the lenticular development of coarser-grained sandstones, grits, and conglomerates. These modify the simple or ideal cyclothems in the Brenot and Dresser Formations, but are less prevalent in the Gething. Such details of sedimentation are better appreciated in closely drilled sections (*see* McKechnie, 1955; Hughes, 1964). Clastic sediments of the Crassier Group were derived from westerly sources. The formations thicken from east to west. The sandstones contain much clastic chert, indicating that extensive areas of sedimentary rocks in the western hinterland were uplifted and eroded in Crassier times. Some of the eroded sedimentary rocks were of Triassic age, from the evidence by McKechnie (1955). Grains and larger clasts of sedimentary rocks, shales, siltstones, and ironstone, in the Crassier sandstones, mostly resulted from penecontemporaneous erosion of the coal measures.

Dating of the Crassier Group is incomplete due to the scarcity of good fossil floras in its lower part and perhaps in some measure to uncertainty about the exact range of the Gething flora. The Crassier Group lies between the Monach Formation (Beaudette Group) and the Moosebar Formation (Fort St. John Group), a considerable interval of time represented by the Hauterivian (and possibly uppermost Valanginian), Barremian, and Aptian, and a lower part of the Albian Stage, in terms of the international standard stages of Muller and Schenck (1943). The Moosebar Formation can be assigned to the Middle Albian, from reports of its marine faunas by Stelck *et al.* (1956), and Stott (1963), but Jeletzky (1964) notes a Lower Albian range (or its possibility) to the *Lemuroceras-Beaudanticeras affine* fauna, which occurs in the Moosebar Formation outside the Pine Valley (McLearn and Kindle, 1950). Collections from the Brenot Formation, in the lower part of the Crassier Group, yielded one identifiable species, *Pityophyllum nordenskiöldi* (Heer) Kryshtofovich, identification by McGregor (1961). This only allows a general Lower Cretaceous dating (Neocomian-Aptian-Lower and Middle Albian), from Bell (1956), and from the report of Stott (1962), as the species is common to the Kootenay and the Luscar-Gething flora. The Luscar-Gething flora, represented by four species in one collection from the Dresser Formation, and more fully in collections from the Gething Formation, can be dated Barremian?-Aptian-Lower and Middle Albian. Bell (1956) assigned the Luscar-Gething flora to the Aptian, but noted its possible lower range into the Barremian; a later report by Stott (1963) places its upper range in the Middle Albian. Collections from the undivided Crassier Group by Mathews (1947) were referred to the Aptian by Bell. Earlier floras, in particular the Kootenay flora of Neocomian-Barremian age (Bell, 1956), have not been identified in collections from the Crassier Group in previous reports nor found in this work.

In general, there are three possible stratigraphic relationships for the Crassier Group. (1) The strata were laid down in continuous order, and can be dated latest Valanginian to Middle Albian. (2) The Crassier strata form a continuous succession in the latter part of this interval and are to be dated Barremian?-Aptian to Middle Albian. (3) The Crassier Group includes one or more diastems, or disconformities, which separate an upper sequence of Barremian?-Aptian to Middle Albian age in the Dresser and Gething Formations from an underlying sequence(s) not well dated at present. Such breaks in sequence (3) remain to be proven, and are unrecorded by fossil evidence. In the field, the Crassier coal measures can be mapped as a continuous succession. The formational boundaries represent gradations and mark changes in the pattern of sedimentation; they are thought to be diachronous. Sedimentary breaks are noted in the coal measures, but they are regarded as local (washouts in some cases) and of penecontemporaneous origin. The second possibility implies a disconformity at the base of the Crassier Group, for which field work offers little evidence. Beaudette and Crassier beds lie parallel, but their contact is concealed in most places. The development of the Monach quartzites is variable. Where they form single beds or units, Crassier beds follow them directly. In a few places, where the Monach quartzites occur as several beds separated by concealed intervals, the separation of the Beaudette and Crassier Groups also holds good—coal measures everywhere succeed the uppermost quartzite bed of the Monach. It is possible that the concealed intervals include marine or non-marine beds of some intermediate age (Beaudette to Crassier). The question of a disconformity at the base of the Crassier Group may be resolved by zonations and regional studies.

Brenot Formation.—The Brenot Formation is named from the type section in the French Petroleum Company-Richfield Brenot Creek No. 1 well near Hudson Hope (Hughes, 1964). It can be traced to outcrops farther west in the Foothills. The formation consists of lean coal measures with thin coals and barren cyclothems, and is 750 feet thick around Fisher Creek in the Pine Valley.

The Brenot strata include dark-grey shales; interbedded shales, siltstones, and sandstones; argillaceous siltstones and sandstones and arenaceous mudstones (lithified seat earths); grey or black soft mudstones (lithified underclays); coals; and black, fissile, carbonaceous shales (roof shales). They were laid down in thin beds, following the cyclic order listed above. Cyclothems are thin, 2 to 8 feet; many are incomplete with coals, seat earths, and underclays often missing. Coal seams rarely exceed 18 inches in thickness. Sandstones with minor, thin shales and siltstones, in intervals of 10 to 25 feet, modify this cyclic order; the sandstones are very fine to medium grained, and some have a speckled appearance, due to abundant dark-coloured chert grains. Ripple marking, current bedding, clay galls, inclusions of shale and mudstone, and clay ironstone pebbles are other features of these sandstones. The formation contains much comminuted plant debris; fossil leaves seem rare. In the Peace River area, *Pityophyllum* cf. *P. nordenskiöldi* (Heer) Kryshtofovich indicates a Lower Cretaceous (Neocomian-Aptian-Albian) age for the Brenot Formation.

Thin bedding, and the less perfect nature of the cyclothems, the thinness of the coal seams, and the barren condition of many cyclothems, serve to distinguish the Brenot Formation. The Dresser Formation is recognized by its thick coarser-grained sandstones; the Gething by its greater proportion of shales, thick coal seams, and thick and well-developed cyclothems.

The Pacific Great Eastern Railway cut 2 miles southwest of Beaudette Creek shows the lower part of the Brenot Formation, more than 240 feet of strata: a coal seam 3 feet thick marks the base and contact with Monach quartzites; it is overlain by 26 feet of sandstones with minor thin interbeds of shale and siltstone, which in turn are followed by thin-bedded coal measures. (Such local development of sandstones near the base of the formation is not unusual.) Partial exposures of the Brenot Formation occur on the east limb of the Bickford anticline and the west limb of the Fisher anticline, and are accessible along the tributaries of Fisher Creek. In the latter structure, a sequence of barren cyclothems, lacking coals or with only few thin coal partings, forms the upper part of the Brenot Formation. Debris of thin rust-coloured sandstone plates is a common indicator of the Brenot outcrop on high ground.

Dresser Formation.—The Dresser Formation was described by Hughes (1964) from a type section at the west end of the Peace River Canyon. The formation consists of coal measures, with medium- to very coarse-grained sandstones, grits, and minor conglomerates. It includes 1,200 feet of beds measured on the west limb of the Fisher anticline in the Pine Valley.

The coal measures are in intervals of varying thickness, usually less than 80 feet. They contain a series of cyclothems of dark-grey shales; interbedded shales, siltstones, and sandstones; sandstones; lithified seat earths and underclays; coals; black carbonaceous fissile shales (roof shales). Cyclothems are 2 to 12 feet thick, and mostly thin bedded. Many lack coals, and others are partially developed in shale-sandstone-shale sequences. Coal seams are thin, less than 30 inches were observed. Sandstones of the coal-measure intervals are very fine to medium grained, thin bedded, or ripple marked, and clean to argillaceous.

Two sedimentary phases are distinct in the Dresser Formation—the cyclic deposits of the coal-measure intervals, and the coarser clastics which lack regular or simple cyclic order. The latter are represented by the medium- to very coarse-grained sandstones, grits, and conglomerate. They form about one-third of the formation, but are found in greater proportion farther east in the Foothills. The sandstones of both sedimentary phases together make about one-half to two-thirds of the formation.

The coarser clastics were laid down in thick beds, in intervals of 10 to 40 feet. They overlie shales, sandstones, or coals of the coal-measure cycles with scoured and eroded contacts. Their base is marked by shale and mudstone phenoclasts and inclusions, coalified wood fragments, and lenses of pebbles. The coarser clastics are followed by sandstones or shales of the coal-measure intervals in most cases. The medium- to very coarse-grained sandstones are current bedded, and in places contain lenses of grits and conglomerates with pebbles of chert and lesser quartzite. Quartz and chert are the major clastic components; abundant grey, brown, and dark-coloured chert grains give a speckled appearance to the coarser-grained sandstones. There are few beds of conglomerate. They have clasts of chert, mostly grey, brown, and dark coloured, and lesser amounts of quartzite. The clasts are well rounded, with median diameters one-quarter to 1 inch commonly; the cement is siliceous. Beds and intervals of the medium- to very coarse-grained sandstones, grits, and conglomerates are known to be discontinuous or lenticular in the Dresser Formation. This condition is probably general, though not easily demonstrated from exposures in the map-area.

The boundaries of the Dresser Formation are drawn to include the major development of medium- to very coarse-grained sandstones, grits, and conglomerates. This separates the formation from the rest of the Crassier coal measures, and excludes transitions which are present, though not of mappable order. The coarser clastics make thin interbeds in places in the upper part of the Brenot Formation. They persist into the lower third or half of the Gething Formation, where they become more dispersed and widely separated.

Evidence for dating the Dresser Formation, Lower Cretaceous (Barremian?-Aptian to Middle Albian), comes from representatives of the Luscar-Gething flora, found in the upper part of the formation at Grant Knob in the Peace River Canyon. This flora, *Elatides curvifolia* (Dunker) Nathorst, *Elatides splendida* Bell, *Pterophyllum plicatum* Bell, and *?Pterophyllum rectangulare* Bell, was identified by Mc-

Gregor (1960, 1961); the dating is taken from Bell (1956) and the report of Stott (1963).

Outcrops of the Dresser Formation stand in relief against those of the Gething, though they are not distinctive everywhere. On hilly ground or above tree line, the steeply dipping beds of coarse clastics form ridges. The upper 250 feet of the formation in the core of the Crassier anticline is partly exposed about the river bank and in cuts along the Hart Highway from Crassier to Narod Creeks. Farther west, the Dresser outcrop swings around the southeast plunge of the Fisher anticline. It descends to river-level between Narod and Fisher Creeks, where the beds are much disturbed by folding and faulting. South of the Pine River, the Dresser outcrops about the axis of the Pine River anticline, along Willow Creek. Partial exposures of the Dresser Formation can be found in the unfaulted west limb of the Fisher anticline along tributaries of Fisher Creek, 2 to 5 miles north of the Pine River. Subsidiary folds and faults of the Bickford anticline complicate the Dresser outcrop on the west side of Fisher Creek. In this area, one-quarter mile southwest of Cleveland Creek, the Dresser-Gething boundary is placed at the top of an interval of sandstones, grits, and conglomerates, 45 to 75 feet thick. The conglomerates form thick beds and lenses; they are missing or replaced by sandstones, 1 to 2 miles north of the river plain, and are not found in thrust sectors farther east, around Cleveland Creek.

Gething Formation.—The term Gething is long established, dating from McLearn's first work in the Peace River Foothills (1918, 1923). It denoted productive coal measures in the upper part of the Bullhead succession, and separated them from underlying non-marine sandstones and lean coal measures. In the Pine Valley, the Gething Formation includes 1,600 to 1,800 feet of strata, from Crassier to Fisher Creeks.

Strata of the Gething coal-measure cyclothems, listed in order, are: dark-grey mudstones and dark-grey shales; shales with siltstones and sandstones in partings, and thin interbeds; sandstones, very fine to medium grained; silty, sandy mudstones and argillaceous, silty sandstones, with unbedded texture, coalified plant debris, with or without pyritic concretions (lithified seat earths); grey or black soft mudstones, with plant debris and listric surfaces (lithified underclays); coals; thin, black, fissile, carbonaceous shales, with or without plant debris (the roof shales). The Gething cyclothems are well developed and tend to be complete and regular. They are between 5 and 25 feet thick.

Coal seams are numerous in the Gething Formation. Drilling and exploratory cuts provide the best record of seams. In a report of coal surveys, McKechnie (1955) showed seven seams in the upper 1,000 feet of the Gething Formation to attain a thickness greater than 10 feet (seams include bone and non-coaly partings and beds). Another group of seams is known from the lower 250 feet of the Gething Formation, west of the entry of Cleveland Creek on the river plain of the Pine: three of the seams exceed 5 feet; the thickest, of 18 feet, contains shales and siltstones in beds of 5 to 3 feet (*see* McKechnie, 1955, p. 17). Coal seams vary in thickness from place to place; splits appear to be common, for example, seam 76 of McKechnie (1955) in the Noman anticline east of Fisher Creek. Coals of the Gething Formation are of good quality, but seams are deeply weathered. The coal structure is banded, and canneloid coals are few, as for the Crassier Group generally.

Sandstones of the cyclothems form beds up to 15 feet thick; they are plane bedded, current bedded, or ripple marked; some are argillaceous or carry thin interbeds of shale and layers of clay galls. In places, mudstones and shales follow sandstones and shales in incomplete, or reverse, cyclic order. Medium- to very coarse-grained sandstones with grits and pebbly layers occur in the lower part of the formation, four or perhaps six of these beds in the lower 600 feet, west of Crassier Creek.

The proportion of sandstones of all types decreases upward in the Gething Formation. Inversely, the proportion of shales and mudstones increases and exceeds that of sandstones in many cyclothems. The lower sand/shale ratio, the thicker, more complete, and more regular cyclothems, and the greater frequency and thickness of coal seams distinguish the Gething from the rest of the Crassier Group.

The Gething has few exposures, mostly those in exploratory cuts and stream banks. Its topography is subdued, but less recessive than for the overlying shales and mudstones of the Moosebar Formation. The two formations occupy the valleys of Crassier and Fisher Creeks, and flank the axis of the Pine River anticline along Willow Creek. Maps, sections, and notes of these outcrops by McKechnie (1955) show the many small folds and faults which are numerous in the Gething, and in the Crassier coal measures generally.

The Gething follows the Dresser Formation in continuous order, and the boundary is selected to mark a change in sedimentation. The upper boundary of the Gething is a sharp wave-cut contact with the basal conglomerate of the Moosebar Formation. The flora of the Gething Formation is Lower Cretaceous (Barremian?-Aptian-Lower and Middle Albian) age. Species collected from the Crassier and Fisher Creek areas, and identified by McGregor (1960, 1961), are listed: *Thallites blairmorensis* (Berry) Lundblad; ?*Coniopteris brevifolia* (Fontaine) Bell; *Cladophlebis virginiensis* Fontaine emend. Berry; *Cladophlebis strictinervis* (Fontaine) Bell; *Sphenopteris latiloba* Fontaine; ?*Sphenopteris göpperti* (Dunker) Seward; *Ginkgo pluripartita* (Schimper) Heer; cf. *Pterophyllum rectangulare* Bell; *Elatides splendida* Bell; *Elatides curvifolia* (Dunker) Nathorst; *Pityophyllum* cf. *P. nordenskiöldi* (Heer) Kryshtofovich; *Athrotaxites berryi* Bell; *Podozamites lanceolatus* (Lindley and Hutton) Schimper. A pelecypod fauna, obtained 150 feet below the Moosebar contact at Fisher Creek, contained: *Corbula* sp. indet.; *Mactra*? sp. indet.; *Astarte*? sp. indet.; *Tellina* sp. indet.; *Unio*? sp. indet. In his report of the identifications, Jeletzky (1957) commented on the brackish water or marine aspect of the fauna. The fossils occur in mudstones, and indicate a shallow marine incursion of the containing cyclothem. McKechnie (1955) reported *Monotis subcircularis* as a derived fossil of Triassic age in the Gething Formation.

Undivided Crassier Group.—The Crassier coal measures are treated as a single unit in their western outcrops. The lower 2,250 feet are present in the sides of the Pine Valley, west of the Bickford anticline, and a thicker sequence, estimated at 2,500 to 3,000 feet, 3 miles upstream along Big Boulder Creek. Higher beds were removed by post-Cretaceous erosion.

In comparison with eastern outcrops, more sandstones are seen in the undivided Crassier Group. A general increase of sandstone is noted in groups of cyclothems also, the beds being thicker and the sand/shale ratio being greater in interbeds. Cyclic order is less distinct for some of the undivided Crassier Group. Beds of medium- to very coarse-grained sandstones are numerous; in places they appear dispersed in the section, with no obvious limits of stratigraphic distribution; elsewhere they are thick and closely grouped with sandstones of cyclothems. Differences in Crassier lithology from east to west are not extreme. They are thought to result from gradual changes, so the separation between the divided and undivided Crassier Group is arbitrary in some measure. This may also reflect the limitations of exposures. Again, complicated structures handicap correlating the exposures.

On the west limb of the Bickford anticline, the lower 800 to 900 feet of Crassier beds, in a faulted sequence which corresponds to the Brenot Formation, contains sandstones 5 to 15 feet thick separated by concealed intervals and by interbedded dark-grey shales and mudstones, carbonaceous shales, thin coals, siltstones, and sandstones. At the top of the sequence, a thick conglomerate with pebbles of chert and lesser quartzites corresponds to the base of the Dresser Formation. In the Coyote Creek area, a synclinal complex with several angular folds contains coal measures with medium- to very coarse-grained sandstones and few beds of conglomerate. A fault on the west divides its structure from the Big Boulder anticline. About 900 feet of Crassier beds outcrop on the anticlinal crest. Sandstones of fine to coarse grain are exposed in separate beds, here. They become close bedded and more frequent higher in the sequence. Little else of the lithology is visible. Farther northwest from the Pine Valley, Crassier outcrops continue around the plunge of the anticline. Coal measures with much sandstone of cyclic and non-cyclic order and conglomeratic sandstones are found 3 miles upstream on Big Boulder Creek. To the south and southwest, the 2,250 feet of Crassier beds in the sides of the Pine Valley include coal measures with medium- to very coarse-grained sandstones and few conglomeratic sandstones. Sandstones are common in the middle and upper parts of this sequence, starting about 400 feet above the base. Details of the lithology, the interbeds of mudstone, shale, siltstone, sandstone, and coal, are exposed along streams. There are no obvious stratigraphic divisions, and no consistent marker beds to trace out the stratigraphy. The Crassier beds are in a series of folds, which appear to be disharmonic structures and to be modified by small faults. Crassier beds are mapped as narrow fault slices in the footwall of the Pyramis thrust. The beds are concealed, and mapping depends on their stratigraphic position overlying a quartzite bed above sandstones in the upper part of the Beaudette Group, and the local spoil of carbonaceous and ferruginous siltstones and sandstones with plant debris. In the Pyramis syncline, 3 miles northwest of the Pine Valley, interbeds of dark-grey shales, carbonaceous shales, arenaceous mudstones with plant debris, siltstones, sandstones, and few sandstones with layers of chert and quartzite pebbles and cobbles outcrop at elevations about 4,500 feet. These beds form parts of a sequence of 250 feet, which is referred to the lowermost part of the undivided Crassier Group. Their exposure is small, and their field relation to the Beaudette Group is not evident in the limits of the map-area. It is probably a disturbed or faulted contact on the west. On the east, toward the axis of the Pyramis syncline, the Crassier beds are thought to overlie the Beaudette Group in conformable order.

Mapping gives little information on coal seams in the undivided Crassier Group. Seams observed were less than 4 feet thick, and those sampled were much weathered, according to analyses.

Fort St. John Group

The Fort St. John Group is divided into five formations: the Moosebar and Commotion in the lower part; the Hasler, Goodrich, and Cruiser in the upper part. The Gates, recognized in the eastern area of the Peace River Foothills, is partly equivalent to the Commotion Formation of the Pine Valley. The division of the Fort St. John Group is based on the alternation of predominantly shale units, the Moosebar, Hasler, and Cruiser, with sandstones and conglomerates composing the Commotion (Gates) and Goodrich Formations.

In the Foothills, the Fort St. John Group was defined by Wickenden and Shaw (1943, p. 3). It has been redescribed by Stott (1961c), who designated type sections of formations, excepting the Moosebar. The Fort St. John Group requires a different classification for the subsurface and outcrop of the Plains. Correlations of the sections in the Foothills and the Plains have been made by Stelck, Wall, Bahan, and Martin (1956, p. 8), and Stott (1961A, p. 5). Seemingly, the different classifications of the Fort St. John Group reflect facies changes and intertonguing of shale and sandstone units.

Comparable changes occur within the Foothills also. The Commotion Formation thins northward from the Pine Valley to the Peace River Canyon. The upper part of the Fort St. John Group contains a large increase in the proportion of sandstones from east to west within the Pine Valley. The sections differ considerably across a distance of 10 to 12 miles, from Commotion Creek to the west escarpment of the Hulcross syncline overlooking Fred Nelson Creek (Fig. 4). The thickness of the Goodrich Formation is doubled. Correspondingly, the shale units, the Hasler and Cruiser Formations, thin westward. Sandstones replace shales to the west, thereby indicating the sediments were derived from westerly sources.

Fort St. John strata are marine, except for parts of the Commotion Formation. Three distinct macrofaunas are widely recognized in the Fort St. John Group, and occur in the following associations (after McLearn; McLearn and Kindle, 1950, pp. 73–96):—

Neogastroplites fauna: Goodrich Formation, the lower "four sandstone" member of the Sikanni Formation (Sikanni Foothills), and the upper part of the Shaftesbury Formation (Peace River Plains).

Gastroplites fauna: Middle and upper marine beds of the Commotion Formation (Members (ii) and (iii) in the Pine River Foothills, the Hulcross and Boulder Creek members of Stott (1963)); and the lower part of the Hasler Formation (Peace River Foothills and Plains).

Lemuroceras and *Beaudanticeras affine* fauna: Moosebar Formation, the Gates Formation (Peace River Foothills and Plains), and (?) the lower marine beds of the Commotion Formation (Member (i) in the Pine River Foothills, the Gates member of Stott (1963)).

The faunas allowed McLearn (1945A), Reeside and Cobban (1960) to assign a general Albian age to most of the Fort St. John Group. Work on microfaunas by Stelck, Wall, Bahan, and Martin (1956), and on the Commotion flora by Bell (1956) confirms this dating. Jeletzky (1964) places the *Lemuroceras-Beaudanticeras affine* fauna in the Lower and Middle Albian, the *Gastroplites* fauna in the Middle and (?)Upper Albian, and the *Neogastroplites* fauna in the Upper Albian.

Part of the Cruiser, the uppermost division of the Fort St. John Group, is likely of Upper Cretaceous age and referrable to the Cenomanian according to the report (Henderson, 1954, p. 2282) of *Acanthoceras* in part association with fish scales from the upper shale member of the Sikanni Formation, and the report (Stelck in Gleddie, 1954, p. 492) of *Pleurohoplites* in association with *Neogastroplites* and fish-scale beds within the Shaftesbury Formation, a part correlative of the Cruiser Formation. Beds with fish scales (the Fish Scale Zone) overlie and extend into the range of the *Neogastroplites* fauna. They provide an approximate index of the Lower-Upper Cretaceous boundary in the Foothills generally, and in the west border of the Interior Plains (Reeside and Cobban, 1960, p. 28). Beds with fish scales are present in the lower third of the Cruiser Formation in the Pine Valley. The Fort St. John Group is overlain by the Dunvegan Formation, for which a Cenomanian age is generally accepted.

Moosebar Formation.—The Moosebar Formation was named by McLearn (1923, p. 5), from the Peace River Canyon. Wickenden and Shaw (1943) then used the term in the Pine Valley. It supplanted the name Pine River Shales (of Spieker, 1921), now discarded. The Moosebar has a basal conglomerate of chert and quartzite pebbles, with thickness varying from a few inches to 20 feet. The remainder consists mostly of dark-grey, rubbly, and partly calcareous mudstones and shales, with minor beds of argillaceous sandstones, and ironstone bands. Thin bentonite layers, to one-quarter inch thick, were noted in places by Wickenden and Shaw (1943) and Spivak (1944). The Moosebar Formation is 1,400 feet thick in the Pine Valley, from measurements in the west limb of the Hulcross syncline.

The basal conglomerate of the Moosebar overlies the Gething coal measures, with a clean-cut surface of slight relief, resting on different beds of sandstone, shale, and coal, from Fisher Creek to Beaudette Creek. The contact is regarded as disconformable, and indicating over-all marine transgression. Clasts of the Moosebar conglomerate are 0.1 to 1.0 inch median diameter, mostly chert pebbles, black, dark, and varicoloured, and well rounded. Small rounded cobbles occur in some layers. The cement is siliceous, but sandstone and mudstone matrices are common in the upper part. Locally, conglomeratic layers or sandstones make a narrow transition to the overlying mudstones and shales.

The Moosebar shales and mudstones are soft and easily weathered. Their outcrops in the valleys of Fisher, Beaudette, Falls, and Crassier Creeks are thickly forested. Exposures are in cutbanks of the creeks. They show features which distinguish the Moosebar from other shale formations of the Fort St. John: a smaller proportion of thin interbeds and partings of sandstones; the development of argillaceous sandstones in beds, usually 1 to 6 feet thick; the indistinct bedding and curved slabby weathering surfaces of the mudstones; the extensive bands of ironstone, and in places segmented, fusiform ironstone concretions perpendicular to the banding.

The lower part of the Moosebar, about 500 feet of beds above the base, is exposed in folded and faulted ground along the lower course of Beaudette Creek. Here, shales and mudstones with ironstone bands and concretions, and several beds of argillaceous glauconitic sandstones, overlie the basal conglomerate of few inches thickness. A cutbank in Crassier Creek, 2.75 miles upstream from the Hart Highway, exposes most of the lower 600 feet of the Moosebar Formation in a section described by Wickenden and Shaw (1943, p. 4). The British Columbia Government Pine River No. 1 well, at Commotion Creek, cuts the Moosebar Formation in the interval 1,081 to 2,424 feet depth, but the Moosebar-Commotion boundary is not clear. The combined Moosebar-Commotion interval at Commotion Creek, calculated as 2,850 feet in the drilled and surface sections, compares with 2,825 feet mapped in outcrop near Browns Creek, 12 miles to the west. The record of drill cuttings (Appendix 3) can be taken as representative of the Moosebar section, without including any large error due to minor disharmonic folds and faults.

The Moosebar Formation contains representatives of the *Lemuroceras-Beaudanticeras affine* fauna in collections from the Peace River (*see* Beach and Spivak, 1944; McLearn in McLearn and Kindle, 1950). The formation can be dated Middle Albian, after Jeletzky in Stott (1963), and after Stelck, Wall, Bahan, and Martin's (1956) dating by foraminifera. Macrofaunas from the Moosebar in the Foothills area of the Pine Valley contain pelecypods of less stratigraphic value at present. These include: *Yoldia* cf. *Y. kissoumi*, and species of the genera *Pecten (Entolium), Corbula, Goniomya, Modiolus*?, *Protocardium*?, *Psilomya*, and *Lima*, listed by Spivak (1944); *Pecten* sp. noted by Mathews (1947); *Pecten (Camptonectes?)* sp. indet., and *Lima* sp. indet., in the writer's collections, identified by Jeletzky (1956).

Commotion Formation.—The term Commotion, introduced by Wickenden and Shaw (1943, p. 5), denotes an assemblage of sandstones, shales, and conglomerate, followed by thin coal measures. The term first applied to the Pine Valley. Commotion strata are now widely recognized in the Foothills, from the Peace River southeastward to the Kakwa River (Mathews, 1946; Stott, 1963). The Commotion occupies the same stratigraphic position as the Gates Formation, above the Moosebar and below the Hasler shales. The Gates Formation of McLearn (1923) outcrops in the east parts of the Peace River Foothills and the nearby Plains. It is equivalent to the lower part of the Commotion Formation, from the correlation by

Wickenden and Shaw (1943), and its restatement by McLearn and Kindle (1950), and Stott (1963). The upper part of the Commotion then corresponds to the lower part of the Hasler Formation on the Peace River—the two parts containing representatives of the *Gastroplites* fauna (Beach and Spivak, 1944). The Hasler represents a longer interval of shale sedimentation north of the Pine River.

It is practicable to divide the Commotion Formation into four members in the map-area. Outcrops in the west limb of the Hulcross syncline, in the escarpment from Crassier and Fred Nelson Creeks to Browns Creek on the southeast, provide the entire sequence as noted in the following summary.

Member (i): The lower beds, above the Moosebar shales, are flaggy greenish-grey sandstones overlain with a chert pebble conglomerate; thickness, 600 feet.

Member (ii): The overlying member contains hackly greenish-brown siltstones and silty mudstones, dark-grey shales and carbonaceous shales, few thin sandstone beds; thickness, 350 feet, largely concealed.

Member (iii): The third member consists of platy and flaggy sandstones followed by 40 to 120 feet of conglomerate and conglomeratic sandstones; thickness, 250 feet.

Member (iv): The uppermost member is a coal-measure sequence, with shales, mudstones, siltstones, sandstones, and thin coals; thickness, 225 feet.

From regional studies, Stott (1963) named three members for the Commotion Formation. They correspond, in part and in general terms, to the divisions given here: the Gates Member (Stott, 1963), an equivalent of the Gates Formation, to Member (i); the Hulcross Member, to Member (ii); the Boulder Creek Member, to Members (iii) and (iv) together. Facies changes in cycles of marine regression and transgression with non-marine intervals seem characteristic of the Commotion, as illustrated by its relation to the Gates Formation, and by sections from the Pine River to the Kakwa River given by Stott (1963).

Exposures do not yield a complete record of the Commotion Formation in the west limb of the Hulcross syncline. Member (i) forms an outlier at Falls Mountain to the west. The Commotion anticline contains the easterly outcrops of Members (iii) and (iv), at Commotion and Goodrich Creeks.

Member (i): About 575 feet of strata at Falls Mountain are assigned to Member (i). The lower boundary with the Moosebar lies above a brief transition of thin-bedded shales, siltstones, and sandstones.

The lower 425 feet of Member (i) consists of sandstones, in beds 5 to 20 feet thick, separated by shaly interbeds, and intervals of thin-bedded sandstones, shales, siltstones, and sandy mudstones. Beds of sandstones are grouped in three major associations, each thicker than the lower, the uppermost 80 to 120 feet thick. The sandstones are current, platy, or flaggy bedded, and ripple marked; they contain clay galls and pebble layers, also much muscovite, and much fine plant debris on bedding planes. Most of the sandstones weather in green and buff colours.

In the overlying 125 feet of Member (i), there are concealed parts and exposures of interbedded shales and sandstones followed by dark-grey shales, black carbonaceous shales, arenaceous shales, and mudstones with few coaly partings. Sandstones with indistinct casts of pelecypods, in a bed 2 to 3 feet thick, mark the top of this interval. Flaggy sandstones and rubbly brownish siltstones and mudstones complete the rest of the section.

The Falls Mountain section of Member (i) records marine advances and retreats, ending in non-marine estuarine deposition. Farther southeast in the Foothills, Member (i) contains non-marine phases and coal measures, from which Stott (1962, 1963) reported the Luscar-Gething flora.

Member (ii): Very little is seen of this unit from Browns Creek to Crassier Creek. It is presumed to be marine, but may include non-marine parts. In addition, thin interbedded shales, siltstones, and sandstones, together with a concealed interval between 2,010 and 2,150 feet elevation in the Commotion anticline, may be doubtfully referred to Member (ii) (*see* following description).

Member (iii): In the eastern outcrop of the Commotion Formation, there are 400 to 450 feet of beds in the valley sides at Commotion Creek, above the drift level at 2,020 feet elevation. Here, shales and silty shales with thin interbedded sandstones mark the base of the section along the Hart Highway. The following concealed interval of 140 feet is overlain by beds which belong to Member (iii). These consist of flaggy medium- to coarse-grained sandstones with conglomeratic layers in the lower 70 to 80 feet, and a massive conglomerate 40 to 50 feet thick, above. Coal measures of Member (iv) overlie the conglomerate.

A change upward from fine to coarse clastics occurs in Member (iii) within the Pine Valley. The conglomerate varies in thickness, and is 30 to 120 feet thick in the western outcrop. It is composed of chert pebbles and fewer quartzite pebbles strongly bound by siliceous cement. These clasts are well rounded, with median diameter usually less than three-quarters of an inch; the cherts are varicoloured, brown, black-grey, red, and green. Grits, with speckled appearance due to light- and dark-coloured granules of chert, occur at the base of the conglomerate in places. Member (iii) serves a useful reference horizon in mapping, being resistant to erosion and well exposed. It lines the crest of the escarpment formed on the Commotion beds in the west limb of the Hulcross syncline. A traverse along the escarpment, from the Moberly River to the Pine River, shows that conglomerates, grits, and conglomeratic sandstones make lenticular bedded units in Member (iii), and that these overlie each other, distributed in echelon, progressively higher in the sequence from north-northwest to south-southeast; the uppermost unit of conglomerate appears to be continuous. The field relations, if confirmed by detailed mapping, suggest a marine regression from north-northwest to south-southeast, or about this direction. Also, depending on the correlation of Member (iii) with lower marine shales in the Hasler Formation at the Peace River, they suggest a possible stratigraphic break of local or regional extent at levels corresponding to the boundary between Members (iii) and (iv) of the Commotion Formation. There is no record for such break in the Fort St. John Group in the Foothills, and detailed zonations are lacking.

Dating of Member (iii) depends on few fossil records from the Pine Valley. Stelck (1942) reported *Inoceramus cadottensis* McLearn and *Gastroplites* sp. cf. *G. kingi* McLearn, and referred these to the *Gastroplites* fauna; Wickenden and Shaw (1943) also reported *Inoceramus cadottensis*, a member or associate of the *Gastroplites* fauna, from " sandstone beds just below the conglomerate." The *Gastroplites* fauna can be dated Middle Albian with a possible range into the Upper Albian (following Jeletzky in Stott, 1963; Jeletzky, 1964). According to Beach and Spivak (1944), the *Gastroplites* fauna occurs about 600 feet above the base of the Hasler Formation, near Starfish Creek in the Peace River Canyon (*see* also McLearn, 1945, p. 10; McLearn in McLearn and Kindle, 1950, p. 80). The containing beds are shales, siltstones, and thin-bedded sandstones, a distinct arenaceous phase of Hasler deposition. They can be taken as correlatives of Member (iii), which undergoes lateral facies change (*see* Wickenden and Shaw, 1943). Also the correlation may extend to part of the interval, Members (ii), (iii), and (iv) of the Commotion, and to an upper range into the Hasler Formation, in the Pine Valley.

Member (iv): The coal measures of Member (iv) are 160 to 180 feet thick at Commotion Creek. They contain black, fissile, carbonaceous shales, with layers of plant debris; dark-grey shales and dark-grey rubbly mudstones; sandstones and siltstones, laminated and thinly interbedded with shales; fine- to coarse-grained sandstones with pebbly layers in places (pebbles of chert, fewer quartzites and clay ironstone); dark-grey and brownish arenaceous mudstones; and unbedded argillaceous sandstones with coarse plant debris and coalified traces of " rootlets " (lithified seat earths); grey and dark-grey soft mudstones (lithified underclays—not always present); and coals, with banded structure. The beds follow a cyclic order as listed. Most cyclothems show regular development and are 3 to 15 feet thick. In 100 feet of the exposures at Commotion Creek, seven cyclothems are complete with coal seams or partings; two cyclothems contain closely paired seams, their lower seams 53 and 116 feet above the base of Member (iv) respectively. Coal occurs in thin seams, 3 to 18 inches thick or as partings; the thickest seam is 22 inches. Sandstones and sandstones with thin interbeds of shale and siltstone are 1 to 8 feet thick.

The lowest beds of Member (iv) are concealed in the 6-foot interval above the conglomerate of Member (iii). This contact seems abrupt and without interbedding or gradations. A 10-foot band of sandstones with conglomeratic lenses and argillaceous sandstones lies at the top of Member (iv). The sandstones are in several beds and lenses with thin shale partings and layers; parts are strongly indurated and concretionary. They contain coarse plant debris, rootlets, and fossil leaves. All exposures of Member (iv) show coal measures as described; marine beds are absent or restricted to included phases of the cyclothems. Most of Member (iv) is exposed at Commotion Creek. Other exposures at Goodrich Creek and at Fred Nelson Creek in the western outcrop are incomplete.

Coals of Member (iv) record former subaerial, terrestial environments. A flora from the top sandstone band of Member (iv), obtained *in situ* and in locally derived float, contains fragments of *Menispermites reniformis* Dawson; cf. *Pseudocycas dunkeriana* (Göppert) Florin; cf. *Protophyllum;* angiosperm leaf, indet. Also, cf. *Laurophyllum* comes from 100 to 120 feet above the base of Member (iv). McGregor (1960) identified these in the writer's collections from Commotion Creek. A previous collection from Commotion Creek, listed by Bell (1956), contained *Pterophyllum validum*? Hollick; *Ficus*? *fontainii*? Berry; *Menispermites reniformis* Dawson; *Menispermites potomacensis* Berry; *Fontainea grandiflora* Newberry. Bell dated the collection Albian-Cenomanian, with a tentative note "a late Albian age for the plant-bearing Commotion beds (Member (iv)) is . . . quite possible." The overlying Hasler beds are undated by fossils. The succeding Goodrich beds contain the *Neogastroplites* fauna. Consequently, marine faunas allow a Middle-Upper Albian dating for Member (iv).

Hasler Formation.—The Hasler Formation was named from the Foothills area of the Pine Valley by Wickenden and Shaw (1943). It consists of dark-grey and black shales; hackly, silty and sandy shales, and mudstones, with thin interbeds (less than 3 inches), laminæ, and partings of grey sandstones; and commonly thin ironstone banding. In eastern outcrops the Hasler Formation is 1,000 to 1,100 feet thick, and is mostly argillaceous. West of Mount Hulcross, the 825 feet of beds assigned to the Hasler include much sandstone. Shales with thin pebble layers mark the base of the formation.

The Hasler outcrops in the Commotion anticline. In the Hulcross syncline, where it borders the river plain of the Pine, the Hasler-Goodrich boundary descends to 2,350 feet elevation about the synclinal axis. There are many exposures in cutbanks along tributary streams from Fur Thief Creek to Goodrich Creek. Red staining and iron rust coatings, and yellow sulphur efflorescence, in places associated with gypsum, are features of cutbanks and talus rubble. They suggest alteration of pyrite; finely disseminated pyrite occurs within the shales and mudstones.

The lower boundary of the Hasler Formation shows an abrupt marine transgression. At Commotion Creek, shales, mudstones, with siltstone and sandstone layers, and thin pebble layers (1 to 6 inches thick) of the lower 8 feet of the Hasler, lie above the sandstone band marking the top of the coal measures (Member (iv)) of the Commotion Formation; the detail of the contact is concealed. The overlying shales represent marine sedimentation which persisted for the remainder of Fort St. John time. A bed of sandstone, grit, and conglomerate occurs in both west and east sections, at corresponding stratigraphic levels, 560 to 530 feet respectively above the Commotion-Hasler boundary. It is 7 to 8 feet thick, base unseen, near Fred Nelson Creek, and 10 inches thick, with eroded wave-cut base, at Commotion Creek. The bed forms a horizon separating two divisions of the Hasler in the western section: a lower, mainly argillaceous division, and an upper, more arenaceous facies of the Hasler (Fig. 4). A transition of interbedded shales and sandstones relates the Hasler to the Goodrich Formation. The mapped boundary includes the major developments of sandstone in the Goodrich. It is diachronous, cutting older, lower, stratigraphic levels from east to west. As a result, the Hasler Formation is thicker in the eastern outcrops.

The Gates and Commotion strata fix the base of the Hasler Formation. The overlying Goodrich sandstones separate the Hasler and the Cruiser, which are predominantly argillaceous formations. The extent of the Goodrich is limited eastward. It divides into several sandstone units, and loses definition in the subsurface of the Outer Foothills (*see* following discussion). Where the sandstones disappear, the Hasler and Cruiser shales are then brought under the name Shaftesbury Formation, as in the Plains.

The Hasler Formation contains the *Gastroplites* fauna, 600 feet above the base, in the Outer Foothills of the Peace River area (Beach and Spivak, 1944; McLearn, 1945); no faunas have been reported from higher stratigraphic levels here. Shelly faunas are lacking from the Hasler of the Pine Valley. Accordingly, the formation here can only be dated by its stratigraphic position as Middle-Upper Albian. The possibilities of minor or local stratigraphic breaks, corresponding to the marine transgression of the Hasler upon the coal measures of the Commotion Formation (Member (iv)), and to the middle sandstone and conglomerate bed of the Hasler, may be noted.

Goodrich Formation.—The Goodrich Formation was first described from the Pine River Foothills by Wickenden and Shaw (1943, p. 7). The formation consists of fine- to coarse-grained sandstones which weather in light-grey and buff colours, and in weathering exhibit a flaggy and platy habit. Some sandstones are calcareous and some contain *Inoceramus* sp., and locally abundant *Pteria (Oxytoma)* sp. (identifications by Jeletzky, 1957). Interbeds of shale and siltstones with thin sandstones are present in the formation and may attain thicknesses of 20 feet. A conglomerate bed 30 feet thick, passing to a thinner conglomeratic sandstone in the east, occurs in the lower part of the formation in the area of Mount Hulcross in the Pine Valley. The lower contact with the Hasler Formation is variable, either a brief passage from shales to sandstones or a less definite and broader transition of interbedded shales and sandstones.

Figure 4 illustrates the relationship of the Goodrich to the Hasler and Cruiser Formations as seen in the north wall of the Pine Valley. The Goodrich Formation thickens westward from 500 to 1,225 feet within a distance of 12 miles, largely the result of an increase in the number of sandstone beds composing the formation. The incoming sandstones replace shales in the adjacent parts of the Hasler and Cruiser Formations. Such replacement is extreme in the west limb of the outcrop, at the west escarpment of the Hulcross syncline. The boundaries of the Goodrich with the Hasler and Cruiser Formations become indistinct here. They can only be placed with difficulty for lack of exposure, and because interbedding of sandstones, siltstones, and shales form a broad transition above and below the Goodrich Formation. Much of the sandstones appearing near the west limit of the outcrop are argillaceous and silty. Stott (1961c) noted the westward thickening of the Goodrich Formation in sections 5 and 7 miles northwest of the Pine Valley. Different thicknesses assigned to the Hasler, Goodrich, and Cruiser Formations by Stott mostly reflect alternative placings of the Goodrich boundaries from those given here, but may also indicate local variations in sedimentation in this area. (Sections described by Stott are better exposed than those in the Pine Valley.)

The Goodrich Formation is dated by the *Posidonia*? *nahwisi* et var. fauna listed by McLearn (1945, p. 11), and by its association with the *Neogastroplites* fauna. Jeletzky (1964) recognized both faunas to be coeval, as had McLearn (1945), and assigned them to a generalized *Neogastroplites* zone of Upper Albian age. Wickenden and Shaw (1943) reported *Posidonia*? *nahwisi* var. *goodrichensis* from the Goodrich Formation in the Mount Hulcross-Commotion Creek map-area. They noted a species of *Oxytoma* to be useful in identifying the formation (*see also* McLearn, 1945). However, its distribution seems patchy. The writer's collections include: *Pteria (Oxytoma)* cf. *camselli; Pteria (Oxytoma)* sp. indet.; *Donax* sp. indet.; *Inoceramus* sp. indet.; identified by Jeletzky (1957). For other collections from the Goodrich Formation in the Pine Valley, McLearn (McLearn and Kindle, 1950) listed: *Oxytoma pinania* McLearn; *Pleuromya kissoumi* McLearn; *Tancredia stelcki* McLearn; *Pleuromya wickendeni* McLearn; *Lucina*? *goodrichensis* McLearn. A sandstone bed in the lower part of the Goodrich Formation yielded the flora: *Cornus; Viburnum; ?Dalbergites; ?Platanus;* and cf. *Dryophyllum elongatum* Dawson; all in fragments too small for specific identifications. A second lot from the same bed contained: *Protophyllum* cf. *P. leconteanum* Lesquereux; cf. *Laurophyllum insigne* Dawson; *Viburnum* sp.; *Populites dawsoni* Bell. McGregor (1960) identified this flora, and observed that the " relationship to the Upper Blairmore and Dunvegan floras is not clear from this small collection."

Cruiser Formation.—The Cruiser Formation was defined by Wickenden and Shaw (1943, p. 9). The formation is mostly argillaceous, containing dark-grey shales and mudstones, silty shales and mudstones, thin interbeds and partings of dark-grey and light-grey sandstones, and clay ironstone banding. In the west part of the outcrop it includes thicker beds of sandstone. The formation is 450 to 850 feet thick, decreasing westward.

The Cruiser resembles the Hasler Formation. There are few differences in lithology: the Cruiser contains shale-mudstone intervals with smaller proportions of sandy interbeds and partings than is general for the Hasler Formation; shale and sandstone interbeds are more sharply differentiated in much of the Cruiser; and many of its thin sandstone interbeds are less argillaceous or are clean, and some are quartzitic. Also in much of the Cruiser, shale and mudstone interbeds are nonarenaceous or less arenaceous; these intervals are without the hackly appearance common to the Hasler sections. Exposures of both formations reveal the same rubbly weathering and the patchy iron rust coatings with sulphur efflorescence.

The Cruiser Formation thickens eastward from 450 to 800 feet, across the Hulcross syncline. This may result from facies changes, shales replacing sandstones eastward about the formational boundaries, and from greater deposition and preservation of sediments to the east. The lower boundary of the Cruiser appears to be transitional, and it requires a similar definition and selection as for the Hasler-Goodrich boundary. Parts of the Cruiser Formation are missing due to post-Cretaceous erosion; most of the remainder is concealed, but intervals of sandstone, and of interbedded shales, arenaceous shales, and sandstones, become obvious in the west limb of the Hulcross syncline. Near Mount Hulcross the Cruiser Formation includes 5- to 20-foot bands of shales and arenaceous shales with thin interbedded sandstones. These extend eastward to Bowlder Creek, in the lower part of the formation. Low plateau steps mark outcrops of sandstone intervals and bands of the Cruiser, and in places obscure the topographic identity of the formation in the west outcrops. Thin interbedded sandstones, shales, and arenaceous shales, included in the basal Cruiser beds at Bowlder Creek and near the headwaters of Young Creek, also indicate the transition to the Goodrich sandstones.

Exposures of the Cruiser Formation are more complete in the Bissett syncline. One, along an unnamed stream three-quarters of a mile east of Young Creek on the south wall of the Pine Valley, shows the upper 750 feet of beds in a section described by Wickenden and Shaw (1943) and again by Stott (1961). Beds with fish scales are located about 600 feet below the Dunvegan boundary, along a second unnamed stream 1.2 miles east of Young Creek, and also in the lower part of the Cruiser Formation at Young Creek (elevation 2,450 feet). The limits of the Fish Scale Zone remain undefined. The upper 340 feet of the Cruiser Formation is exposed at Stone Creek: shales with siltstone and sandstone in thin flaggy interbeds form the lower 40 feet; in 100 feet of transition, they pass upward to shales and mudstones, and the succeeding 170 feet contains mudstones and lesser arenaceous mudstones, with a few bands of shale-siltstone-sandstone interbeds, and an increase of silty and sandy partings in its upper part; in the uppermost 30 feet of the section, thin interbedded shales, arenaceous mudstones, and argillaceous sandstones underlie beds with *Unio (Pleurobema)* cf. *dowlingi*, which are placed in the Dunvegan Formation; much comminuted plant debris occurs in the upper 200 feet of this section.

The Cruiser Formation can be assigned to the Lower and Upper Cretaceous (Upper Albian-Cenomanian), from its stratigraphic position, and the evidence of the fish-scale beds, without record of any other fossils.

Dunvegan Formation

The Dunvegan strata were first named by Dawson (1881, p. 115) for their extensive outcrop in the eastern Foothills and the Plains. They comprise the youngest formation in the map-area, and are 1,150 feet thick near Stone Creek, and 1,194 feet thick in the section drilled in Sun et al Chetwynd 14-20 at the east limit of the map-area. The lower beds of the Dunvegan consist of interbedded sandstones, siltstones, and shales, commonly bearing *Unio (Pleurobema)* cf. *dowlingi* McLearn. They are 40 to 90 feet thick, and overlie the Cruiser shales in conformable attitude. The main part of the Dunvegan has beds and intervals of sandstones, 8 to 40 feet thick. The sandstones are fine to very coarse grained, with bottom sets and occasional lenses of grits and conglomerates. Intervals ranging to 80 feet thick and separating the sandstones contain thin interbedded green and brown shales, mudstones and siltstones, dark-grey shales and carbonaceous shales, grey siltstones, and micaceous sandstones, together with grey or green sandstones in beds 2 to 4 feet thick. In the Pine Valley, the Dunvegan Formation is without any known marine beds. Plant debris and fossil leaves are common. Minor amounts of coal are recorded in Sun et al Chetwynd 14-20. Muscovite is generally common in Dunvegan sandstones of all types and associations. The green to grey colours of the weathered sandstones are characteristic.

Two informal divisions are given for the Dunvegan, the lower (or *Unio (P.)* cf. *dowlingi*) beds, and the main part of the formation containing thick beds and intervals of sandstones.

Exposures of the lower beds of the Dunvegan show variations in bedding, the proportions of shale and sandstone, concretionary banding, and the abundance of *Unio (P.)* cf. *dowlingi;* this fossil is absent, or not found at some localities. At Stone Creek, the lower beds, about 65 feet thick, overlie the Cruiser with plane separation. They consist of thin-bedded, grey, brown-weathering, micaceous sandstones, shales, and shales with siltstones. *Unio (Pleurobema)* cf. *dowlingi* crowd several bedding planes, and comminuted plant debris is common. Dark-grey shales, 1.5 feet thick, overlie the lower beds, and in turn are overlain by massive green-grey medium- to very coarse-grained sandstone with conglomeratic lenses.

Developments of the lower, *Unio (P.)* cf. *dowlingi*, beds at Caron Creek and Mount Wabi resemble that of Stone Creek. Fossils have not been found in the lower beds at Wildmare Creek. At the localities noted, beds and intervals of fine- to very coarse-grained sandstones belonging to the main part of the Dunvegan rest on the lower beds. The contact is sharp and abrupt, and either wave cut or scoured. West of Commotion Creek there are poor exposures of silty shales with siltstones and thin-bedded sandstones at levels near the lower beds and about the Dunvegan base. They are unfossiliferous. Obscure casts of pelecypods occur in the lowest sandstones of the main part of the Dunvegan in this area.

In the main part of the Dunvegan, sandstone beds and intervals form benches at plateau levels and along the valley sides. Sandstone exposures are less prominent to the east, and less prominent in the upper half of the formation. Sandstone beds and intervals commonly have sharp-cut and eroded basal contacts. In places they contain channelling, cut and fill structures, and steep cross-bedding. In their upper parts they included thin-bedded and ripple-marked finer-grained sandstones. Beds of conglomerate and conglomeratic sandstones occur in the lower 200 feet of the Dunvegan at few localities—between Stone and Bissett Creeks, and along parts of Caron Creek. The lower stretches of Bissett Creek reveal sequences of thin interbedded shales, mudstones, siltstones, and sandstones.

Strata in the upper part of the Fort St. John and in the Dunvegan are generally parallel in the east. Here the Cruiser-Dunvegan boundary represents a change in sedimentation, the incoming of sands, silts, and muds with a non-marine fauna. This shows a quiet marine withdrawal, and there is no evidence for a hiatus at the boundary, that is, at the base of the *Unio (P.)* cf. *dowlingi* beds. West of Mount Hulcross the Cruiser Formation thins and the base of the Dunvegan Formation converges upon horizons in the Fort St. John Group (Fig. 4). The convergence is slight, and details of the Cruiser-Dunvegan boundary are concealed. Available exposures do not establish explanations for the convergence—phases of erosion during late Cruiser times, local uplift and erosion preceding Dunvegan deposition, or replacement of shales by sandstones at the Cruiser-Dunvegan boundary.

The Dunvegan Formation attains its thickest development in the Foothills around the Pine Valley, but it seems unlikely the complete sequence is represented here. Field observations have not warranted subdividing the main part of the formation. The term Sukunka, proposed for non-marine beds in the upper part of the Dunvegan at Dickebusche Creek (Spieker, 1921), has not found use in the Pine Valley. By report, its lithology has little distinction from the main part of the Dunvegan. Another interpretation of the Sukunka Member, " arenaceous fingers (of marine origin) penetrating the overlying Kaskapau shale section " (Stelck, 1962; *see also* Stelck and Wall, 1955), may seem inappropriate from Stott's (1961B) observation at Trapper Mountain.

The Dunvegan Formation contains marine and non-marine strata in the Plains, and in the Foothills southeast of the Pine River. McLearn (1945b), Gleddie (1954), Stott (1961a), and Stelck (1962) list marine and brackish water faunas of the Dunvegan: *Brachydontes multiligera* Meek, *Inoceramus dunveganensis* McLearn, and *Inoceramus rutherfordi* McLearn are characteristic. The faunas are regarded as Upper Cretaceous (Cenomanian) in age by McLearn (1945), and Jeletzky in Stott (1961a). The Kaskapau (Smoky Group) overlies the Dunvegan Formation, and its lower part contains marine faunas, ammonites of the *Dunveganoceras* zone, and microfaunas, of late Cenomanian age (Stelck and Wall, 1955; Stelck, 1962; Stott, 1961a); the Kaskapau Formation ranges into the Turonian stage. *Unio (Pleurobema) dowlingi* McLearn is an index fossil of the Dunvegan, and represents its fresh- or brackish-water facies, in part. In the writer's collections from the lower beds, Jeletzky (1956) identified *Unio (Pleurobema)* cf. *dowlingi, Corbula*? sp. indet., and pelecypods (genus and species indet.).

The Dunvegan flora is well known and has many species; angiosperms make the greater part, from Bell's description (1963). The flora provides independent evidence of a Cenomanian age (ibid.). Collections from the Pine Valley are listed as follows, according to the identifications by McGregor (1960): cf. *Asplenium foersteri* Debey and Ettingshausen; *Elatocladus intermedius* (Hollick) Bell; *Laurophyllum* sp.; *Magnolia* cf. *M. speciosa* Heer (=*M. hollicki* Berry?); *Platanus affinis* Lesquereux (=P. *heeri* Lesquereux?); *Platanus* sp.; cf. *Populites platanoides* Hollick; cf. *Populus; Protophyllum* cf. *P. boreale* Dawson; ?*Protophyllum; Pseudoctenis latipennis* (Heer) Seward; ?*Pseudocycas unjiga* (Dawson) Bell; ?*Pseudoprotophyllum;* cf. *Sassafras (Araliopsis)* spp. in Lesquereux 1874; *Sphenopteris* (*Onychiopsis*?) *psilotoides* (Stokes and Webb) Halle (=*Asplenium dicksonianum* Heer); cf. *Viburnum*.

Dunvegan strata represent aluvial and lagunal sediments built in an extensive delta seaward from a land area on the west. Tectonic conditions, the prevalent subsidence, intermittent still stands, and incursions of brackish water probably did not favour long periods of forest growth. Drifted plant remains, leaves, and fragments are common. Mud cracks, ripple markings, carbonized "rootlets" and "root traces," and coarse plant debris in arenaceous mudstones indicate deposition in shallow water, and some plant growth *in situ*. Distributaries brought coarser sands into shallow lagoons and lakes covering the delta flats. The Dunvegan Formation contains coal seams, but none are exposed in the map-area; coal seams occur in the nearby Plains, at East Pine, and at localities noted by Selwyn (1877), Dawson (1881), and Spieker (1921).

STRATIGRAPHY: SUBSURFACE

In the Pine Valley, wells drilled for oil and gas provide further stratigraphic data. Wells entering simple structures provide useful records. In the Outer Foothills they include Sun et al Chetwynd 14-20-77-23, for Cretaceous, Jurassic, and Triassic strata to 9,403 feet total depth; British Columbia Government Pine River No. 1, for Cretaceous strata from 1,081 to 3,000 feet depth; and Anglo Bralsaman Little Prairie No. 1, for Cretaceous strata to 1,830 feet total depth (a record without cores and dip surveys). In the Inner Foothills, wells entered complex structures with steep-dipping and flexured beds, close folding, and faulting. Their records for Cretaceous, Jurassic, Triassic, and Upper Palæozoic strata are largely of illustrative value. The wells of the Inner Foothills include: Hunt Sands Sun Falls c-18-G; Hunt Sands Sun Boulder b-74-D; Triad Bush Mountain b-23-A (1); and Triad B.P. Bush Mountain a-15-A. Cores and dip surveys are insufficient for full stratigraphic study of their sections, and knowledge of adjacent outcrops is required to interpret the structures therein. Notes on the subsurface geology of wells drilled for oil and gas in the Pine area are set out in Chapter V.

The stratigraphy of the well sections is largely concerned with changes across the Foothills from west to east: changes in formations, their thickness, and sediments. The eastern wells are noteworthy in this regard. Sun et al Chetwynd 14-20 is most important (*see* description in Appendix 4). Its situation on the border of the Foothills and Plains makes it useful in correlating strata in outcrops with those of the subsurface of the Plains. Wells entering complex structures are mostly in the Inner (or Western) Foothills, and do not merit full descriptions of stratigraphy.

SCHOOLER CREEK GROUP

Grey Beds: Halfway, Charlie Lake, and Baldonnel Formations

Formations of the Grey Beds extend under the surface of the Foothills. Part of the Charlie Lake and the complete Baldonnel Formation were drilled to the east of their outcrops.

Sun et al Chetwynd 14-20 penetrated 588 feet of Grey Beds to total depth.* This interval is largely a dolomite sequence. In its lower part, an assemblage of dolomites and shales, with lesser anhydrites, and siltstones with few sandstones, 163 feet thick is placed in the Charlie Lake Formation. The presence of anhydrites with anhydritic clastics and dolomites forms the basis of this assignment. The remainder of the assemblage differs little from the rest of the Grey Beds, locally. The Baldonnel Formation, in the overlying 425 feet, contains light- and dark-coloured dolomites, light-coloured drusy and sub-oolitic dolomites, with dark to black shales, pyritic shales, bedded cherts, siltstones, and minor sandstones. Dolomites compose the major part of this assemblage. The light- and dark-coloured dolomites include crypto to fine crystalline types and recrystallized types. They are commonly silty to varying degrees. More clastics and the appearance of limestones in the upper 35 feet may mark a change to Pardonet lithology.

The Baldonnel Formation in folded, faulted, and incomplete sections in Hunt Sands Sun Falls c-18-G contains limestones, dolomites, and dolomitized limestones, lesser arenaceous carbonates, siltstones, and sandstones with carbonate matrices, and few quartzites.

The wells Triad Bush Mountain b-23-A(1) and Triad B.P. Bush Mountain a-15-A entered disturbed incomplete sections of the Grey Beds which are valuable in proving the stratigraphy of the nearby outcrops. For this reason the well sections are best described with those of the outcrops (*see* previously, Chapter III). The Triad wells reveal the westward extent of evaporitic deposits in the Charlie Lake and thinning of the formation westwards from the Plains.

Baldonnel strata are calcareous in outcrop and dolomitic under the Plains. The lateral passage from limestones to dolomites occurs under the Foothills, and is partly or largely due to regional dolomitization. This change is advanced in Hunt Sands Sun Falls c-18-G, and is virtually complete further east in Sun et al Chetwynd 14-20.

Pardonet Formation

The Pardonet Formation is missing from the subsurface of the Plains in most of northeastern British Columbia. Its absence may result from non-deposition, change of facies and replacement by near-shore sediments, and from pre-Jurassic erosion. Also regional dolomitization may mask the former extent of the Pardonet sediments. Widespread pre-Jurassic erosion can be demonstrated, for under the Plains the Fernie oversteps Triassic to rest directly on Mississippian beds. Subsurface sections of the Pine Valley illustrate an eastward thinning of the Pardonet Formation.

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* Stratigraphic boundaries are found from drill cuttings, which are sampled every 10 feet depth. The boundaries are reported to the nearest 10 feet for transitions between formations and where marker beds are absent or lack obvious register on geophysical logs. Otherwise, boundaries are obtainable to the nearest foot with the aid of geophysical logs and drilling reports. The drilled intervals of formations and groups are listed under "Wells Drilled for Oil and Gas," Chapter V.

In Sun et al Chetwynd 14-20, the Pardonet Formation is absent, unless represented by shales, siltstones, sandstones, limestones, and dolomites in the uppermost 35 feet of the Triassic section.

In Hunt Sands Sun Falls c-18-G, an interval of argillaceous to silty, fine crystalline limestones, and calcareous shales of dark-grey colours, present in faulted and repeated order, is referred to the Pardonet on evidence of lithology and stratigraphic position. In the lower part of the Pardonet cleaner recrystallized limestones, silty limestones, and dolomitized limestones indicate a transition to the underlying Baldonnel Formation. The uppermost beds, with low shale content and low radioactivity, make a convenient marker for the Baldonnel-Pardonet boundary within the transition. No reliable estimates of thickness are at hand; vertical intercepts of the Pardonet Formation are 242 and 243 feet.

FERNIE GROUP

In Sun et al Chetwynd 14-20, the Fernie section, amounting to 507 feet, is complete in normal and unfaulted order. Hunt Sands Sun Falls c-18-G cuts folded and faulted partial sections of the Fernie group over long vertical intervals.

Nordegg Beds

In the subsurface of the Plains, the Nordegg unit is indicated and usually defined from geophysical logs by several characteristics: its low or moderate self-potential; high resistivity, marked off by a sharp deflection of values from that of the overlying Fernie beds; and bands of strong radioactivity. The Nordegg of outcrop and subsurface sections have similar lithology. However, variations in the content of shale and the use of different stratigraphic criteria to define the unit bring about uncertainties in correlating the upper part from outcrop to subsurface.

The same observations hold good in the Pine Valley. Here, in Sun et al Chetwynd 14-20 and Hunt Sands Sun Falls c-18-G, the Nordegg interval, defined according to geophysical logs as in the subsurface of the Plains, contains the following parts:—

Upper Beds: Mostly black non-calcareous shales; few argillaceous limestones and calcareous shales as for the lower beds; siltstones and very fine-grained sandstones in minor amounts.

Lower Beds: Black non-calcareous shales; black argillaceous limestones and calcareous shales; similar limestones and calcareous shales with insets and intergrowths of clear calcite; few siltstones and very fine-grained sandstones; minor chert; few cuttings of phosphatic limestones.

Non-calcareous shales are common throughout the Nordegg interval and predominate in cuttings from the upper beds. They have little distinction from the shales of the rest of the Fernie Group. A large proportion of non-calcareous shales of black sooty appearance is characteristic of the Nordegg, and gives the first lithological index of this unit in drill cuttings; calcareous shales tend to be uncommon in overlying beds. Both lower and upper beds of the Nordegg have bands of high radioactivity.

The Nordegg interval covers 97 feet in Sun et al Chetwynd 14-20, the lower beds 33 feet and the upper beds 64 feet thick.

In Hunt Sands Sun Falls c-18-G, the Nordegg has a vertical intercept of 283 feet, uncomplicated by faulting.

Middle Shales

The term Middle Shales describes an argillaceous unit 313 feet thick in Sun et al Chetwynd 14-20; the stratigraphic thickness in Hunt Sands Sun Falls c-18-G is unknown due to folding and faulting.

Dark-grey and black shales with lesser silty shales compose the most part of this unit; siltstones and sandstones are minor components. The Middle Shales appear to lack any features which may allow its subdivision. Drill cuttings and gamma ray logs show a rather uniform lithology.

Transition Beds

In Sun et al Chetwynd 14-20, the Transition Beds consist of 97 feet of interbedded shales, siltstones, and sandstones. The unit is increasingly argillaceous downward to the Middle Shales. The base of a sandstone bed, 8 feet thick, marks the Fernie-Beaudette boundary locally.

The Transition Beds are absent in Hunt Sands Sun Falls c-18-G probably as a result of faulting.

BEAUDETTE GROUP (UNDIVIDED)

Hunt Sands Sun Boulder b-74-D cuts a long interval of steeply dipping Monteith beds (*see* Chapter V). In Sun et al Chetwynd 14-20, 688 feet of beds is assigned to the undivided Beaudette Group. The latter section is of concern here.

In Sun et al Chetwynd 14-20, the undivided Beaudette Group consists of a uniform sequence of interbedded sandstones, quartzitic sandstones, quartzites, siltstones, and shales. These lack any special distribution or grouping, and the formations of the outcrop are unrecognized. The Beaudette sequence resembles the Beattie Peaks Formation in its sand/shale ratio and bedding, and the Monteith Formation in the nature of its sandstones and quartzites. The sandstones are mostly very fine to medium grained, with much variation in the shale content, the shale occurring interstitially in the matrix, as discrete lithic grains or partly formed around quartz grains and their secondary overgrowths. A range of variation to arenaceous shales appears within the sandstones, and oppositely a range to quartzitic sandstones accompanying growth of authigenic quartz; the quartzitic sandstones contain flecks of shale. Clastic chert is far less common than in the Crassier sandstones. The Beaudette quartzites have much authigenic overgrowths; loose quartz and quartzites showing crystal faces are frequent in the cuttings. The shales are dark grey and black; shales with plant debris and carbonaceous laminæ are present. The coarser clastics form beds up to 12 feet thick, usually, and are separated by interbedded shales, siltstones, and sandstones. Quartzites and quartzitic sandstones occur throughout the section, with some concentrations in the lower and upper third parts.

For the Sun et al Chetwynd 14-20 section, the contrast with the outcrops lies in the eastward thinning and facies changes shown by the distributions of shale, sandstone, and quartzite. It lacks the sandstone development of the Monach. The extent of facies changes and the possibilities of post-Beaudette erosion remain unassessed, and detailed correlations are not available.

CHETWYND BEDS

In Sun et al Chetwynd 14-20, interval 7,375 to 7,620 feet, beds with coal measures, quartzites, and sandstones of marine aspect follow the Beaudette Group and underlie the Crassier coal-measure sequence. Such intermediate beds fall outside the meanings and definitions which apply to Beaudette and Crassier strata in outcrop. Accordingly, they are treated as a separate stratigraphic unit, under the name Chetwynd Beds. A formational definition is inappropriate at present, and under the condition here, namely, that Sun et al Chetwynd 14-20 has the only known section. This section is described in Appendix 4.

The Chetwynd Beds contain coal measures, sandstones, quartzites, quartzitic sandstones, siltstones, and shales. Coal measures are present within the interval 7,500 to 7,575 feet, and evident in core, 7,500 to 7,550 feet. Coals in minor amounts persist in cuttings to 7,450 feet and in a second interval to 7,620 feet depth. The sandstones are very fine to medium grained, with quartz, chert, and lithic grains, mostly of shale. Quartzites and quartzitic sandstones are in less proportion for most of the section. They form a major influx at 7,380 and 7,460 feet, indicating thick beds. The Chetwynd-Brenot boundary is placed at the top of the upper quartzite and quartzitic sandstone. The lower boundary is indistinct, and is taken where coals become insignificant in the cuttings. Many of the Chetwynd sandstones resemble those of the underlying Beaudette Group.

The stratigraphy of the Chetwynd Beds is open to interpretation:—

(1) They may constitute a basal unit of the Crassier Group, affected by oscillations of the former shorelines when coal-measure deposition began.

(2) They may represent a transition between Beaudette and Crassier sedimentation.

(3) If the Beaudette and Crassier Groups are separated in time, the Chetwynd Beds may form an independent stratigraphic unit within the interval late Valanginian-Aptian.

(4) The Chetwynd Beds may include littoral and non-marine facies of the Beaudette. As such, they correspond to the upper parts of the Monach Formation, which indicate the retreat of the Beaudette sea.

(5) Finally, the stratigraphic entity of the Chetwynd Beds may be in doubt from the limited observations. It is possible that they may be resolved into parts of the Beaudette and Crassier Groups in future studies.

CRASSIER GROUP

The Crassier sections, in B.C. Government Pine River No. 1 and Hunt Sands Sun Falls c-18-G, are disturbed by folding and faulting. The section at Sun et al Chetwynd 14-20 is in simple order, and the only one of concern here.

The Crassier coal measures thin eastward under the Foothills. Sun et al Chetwynd 14-20 proves a rapid thinning to 1,488 feet; this accompanies an overall decrease in the grain size of the arenites, compared to the Foothills sections. The Brenot, Dresser, and Gething Formations persist eastward to the Chetwynd section. They are separable on lithological and geophysical logs by the criteria stated for the Foothills outcrops, little modified to account for changes in grain size. The formational divisions at Sun et al Chetwynd 14-20 are then given by the following points:—

(1) Differences in sand/shale ratios. This is manifest by the more arenaceous Dresser Formation.

(2) The distribution of relatively coarser clastics. These occur mostly in the Dresser Formation, and include the range, fine- to very coarse-grained sandstones, grits, and conglomerates (which latter are indicated by grains, granules, and cuttings of chert, and less of quartzite, in the drill cuttings). Most sandstones and coarse clastics of the Dresser are in thicker beds, of 5 to 20 feet. The distribution of thick arenaceous beds allows a direct reading of the Dresser Formation on gamma ray, neutron, sonic, and induction logs.

(3) The incidence of thicker coal seams. Seams are left unmeasured; they are thinner than for the outcrop sections. At Sun et al Chetwynd 14-20, the upper part of the Dresser and the Gething Formation contain many

relatively thick seams; the lower part of the Dresser and the Brenot Formation, thin seams which are progressively less common downward.

Brenot Formation

The 305 feet of beds assigned to the Brenot Formation has increasing sand/shale ratio upwards. Sandstones are mostly very fine grained, but include few grits and conglomeratic layers, showing a change upwards to Dresser sedimentation.

Dresser Formation

The Dresser Formation, 670 feet thick, is distinct in cuttings and logs. Its sedimentary development is like that in the Foothills, but it has less proportions of medium- and coarser-grained sandstones.

Gething Formation

The Gething Formation, 513 feet thick, shows relatively more eastward thinning than the Brenot and the Dresser. This may reflect facies changes, or differences in placing the Dresser-Gething boundary from west to east, or some pre-Moosebar erosion. It is impractical to obtain detailed correlations to examine these factors for the present.

The abundance of coals, the absence of coarser-grained clastics, and thinner bedding of sandstones distinguish the Gething Formation in lithological and geophysical logs at Sun et al Chetwynd 14-20. The Gething section here lacks any obvious or progressive change in the sand/shale ratio. Most of its sandstones are very fine grained. Siltstones are common, and form considerable proportions of much of the drill cuttings.

FORT ST. JOHN GROUP

The wells drilled in the Outer Foothills enter the Fort St. John Group. Sun et al Chetwynd 14-20 has a complete section. It reveals the eastward thinning of all the formations from their nearest outcrops on the west, and facies changes whereby the Goodrich and parts of the Commotion Formation (Member (iv)) lose their identity (Fig. 4). The major change, due to passing out of the Goodrich sandstones, is three separate divisions—the Hasler, Shales with Sandstones, and Cruiser—for the upper part of the Fort St. John Group. Together they correspond in part to the Shaftesbury Formation of the Plains, and also the " Upper Shales " (of Irish, 1958). The facies changes require some modification of nomenclature.

The well sections also give evidence of five glauconite* zones in the Fort St. John Group. These are fully shown in Sun et al Chetwynd 14-20, which permits a reference section, as follows:—

Depth in Drill Cuttings	Formation and Member
(v) 1,840 to 1,860	Shales with Sandstone, fourth (uppermost) arenaceous member: glauconite minor.
(iv) 2,310 to 2,340	Shales with Sandstones, second arenaceous member: glauconite common.
(iii) 3,700 to 3,780	Commotion Formation, Member (ii): glauconite minor.
(ii) 4,610 to 4,700	Commotion Formation, Member (i): glauconite, common at 4,630 to 4,660 feet, and minor for the rest.
(i) 5,650 to 5,890 (base of Fort St. John Group)	Moosebar Formation, lower part: glauconite minor, and widespread throughout, and common at intervals from 5,750 to 5,890 feet.

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* Determinations of glauconite, as listed, are from X-ray diffraction, powder photographs by N. G. Colvin, Department of Mines and Petroleum Resources, Victoria, B.C.

The glauconite is in shales, siltstones, and sandstones, occurring interstitially and in discrete rounded grains. The lowermost zone (i) has its equivalent in outcrop. The other zones are undefined in outcrop, excepting (ii) in the Gates Formation, locally in the Peace River Canyon. For the rest, glauconite is a minor component in parts of the Commotion and Goodrich Formations.

Moosebar Formation

The Moosebar Formation is complete in a section of 1,083 feet in Sun et al Chetwynd 14-20. The formation is contained in the interval 1,081 to 2,420 feet in B.C. Government Pine River No. 1, but its upper boundary here is indistinct or not penetrated.

The formation consists mostly of dark-grey shales, with lesser silty shales, and minor siltstones and sandstones. In addition, its lower part, the lower 400 to 240 feet, from west to east, contains calcareous shales, and glauconitic layers in shales, siltstones, and sandstones. A thin basal conglomerate marks the contact with the Gething coal measures. The upper part of the formation is more arenaceous and shows a change to the Commotion sedimentation. The interval 1,310 to 1,430 feet in B.C. Government Pine River No. 1 includes shales with much siltstones and sandstones, and to assign these and overlying beds to the Moosebar seems doubtful.

Commotion Formation

The Commotion Formation occurs in a section of 1,417 feet at Sun et al Chetwynd 14-20. The section differs from the outcrops in the loss or obscure identity of the uppermost coal measures, Member (iv) of the exposures at Fred Nelson and Commotion Creeks. Consequently, a threefold division is recognized for the formation in the subsurface, namely, Members (1) and (ii), as defined from outcrops, and thirdly a unit designated here as Member (iii and iv) combined.

Member (i): Member (i) is 694 feet thick. It contains, in the lower part, interbedded dark-grey shales, siltstones, and sandstones, followed in the upper 280 feet by sandstones in thick groups of beds with lesser shales and siltstones. Sandstones are mostly very fine to fine grained, and argillaceous to clean, but include coarser-grained sandstones with grit or conglomeratic layers in the upper part. The glauconitic beds, zone (ii), are between 100 and 200 feet above the base of Member (i). Coals form a minor proportion of the drill cuttings, 640 to 650 feet above the assigned base. The Moosebar-Commotion boundary is drawn at the base of a sandstone 10 feet thick, partly an arbitrary marking in a broad transition from the Moosebar shales.

Member (ii): Member (ii) is largely an argillaceous unit 404 feet thick. Most of the drill cuttings are shales, with few silty shales and siltstones; sandstones are in minor proportions, in thin beds. Member (ii) contains a glauconite zone in Sun et al Chetwynd 14-20 (zone (iii) of the Fort St. John Group, already listed).

Member (iii and iv): Marine sandstones of Member (iii and iv) are directly overlain by Hasler shales, in contrast to the stratigraphic relationship observed in outcrops. Sandstones of this unit are very fine to coarse grained, and clean; grits and conglomeratic layers are common. Coals with coal-measure sediments occur in cuttings in minor proportions, within the intervals 130 to 180 and 220 to 260 feet above the base of Member (iii and iv). They are not assumed to be cavings. A question of their correlation with Member (iv) of the Commotion outcrops remains unsettled. As a result, the stratigraphic relation and identity of the uppermost 60 feet of marine sandstones (in Member (iii and iv) of the Chetwynd Sun et al 14-20 section) are in doubt. The combined term (iii and iv), for the upper division of the Commotion in the subsurface, represents a temporary provision for these uncertainties.

Hasler Formation

The complete section of the Hasler Formation is 785 feet thick in Sun et al Chetwynd 14-20. Part of the formation was entered in Anglo Bralsaman Little Prairie No. 1.

In these eastern wells, the Hasler Formation consists of dark-grey shales, with lesser silty shales, and siltstones; sandstones are present in minor proportions, in thin beds.

Shales with Sandstones

The lithological term Shales with Sandstones is proposed for a formational unit of shales with four arenaceous members. This unit occupies a comparable stratigraphic position to the Goodrich, between the Hasler and the Cruiser Formations. The proposed term applies locally to the subsurface in the east margin of the Foothills. Sun et al Chetwynd 14-20 provides a type section in the interval 1,842 to 2,602 feet; and Anglo Bralsaman Little Prairie No. 1, a section in the interval 548 to 1,465 feet.

The Shales with Sandstones form a predominantly argillaceous unit. The arenaceous members are groups of sandstone beds 5 to 25 feet thick, and separated by shales, arenaceous shales, siltstones, and sandstones in thin beds. The sandstones are very fine to coarse grained, variably argillaceous to clean, and with grit or conglomeratic layers in places; the second and fourth arenaceous members have glauconitic parts. The arenaceous members occur as follows: In Sun et al Chetwynd 14-20—first, 2,574 to 2,602; second, 2,315 to 2,358; third, 2,159 to 2,218; fourth, 1,842 to 1,880 feet depth; each member consisting of two distinct sandstone beds: In Anglo Bralsaman Little Prairie No. 1—first, 1,350 to 1,465; second, 1,118 to 1,212; third, 890 to 1,033; fourth, 548 to 627 feet depth; each member containing two or several beds. The intervening shale members consist of dark-grey shales, with lesser silty shales and siltstones; sandstones are minor constituents.

The interval represented by Shales with Sandstones indicates the replacement of Goodrich sandstones by shales to the east. Such facies change is a continuation of that mapped for the upper part of the Fort St. John Group in outcrop in the Hulcross syncline (Fig. 4). Accordingly, the sand bodies of Goodrich deposition together form a wedge, pointed and intertonguing with shales eastward across the Outer Foothills. Sandstone tongues extend out from the wedge and beyond its eastern limit, which lies between Commotion Creek and the Anglo Bralsaman Little Prairie No. 1 well.

The lithological term Shales with Sandstones is regarded as provisional. It may be better replaced by an alternative nomenclature employing separate members of a redefined Shaftesbury Formation, from future knowledge of the subsurface stratigraphy.

Cruiser Formation

In Sun et al Chetwynd 14-20, 632 feet of beds is assigned to the Cruiser Formation. The beds are mostly dark-grey shales, with some silty shales and a few thin sandstones. A sandstone bed, about 12 feet thick, is included in the lower part of the formation. The fish-scale beds are undetected in cuttings. Furthermore, the Fish Scales Marker Horizon (*see* Burk, 1963) is left undesignated. No unique register of the geophysical logs can be ascribed to this horizon, without mapping facies of the Cruiser and Shaftesbury Formations across the west margin of the Plains.

DUNVEGAN FORMATION

The Sun et al Chetwynd 14-20 well, from surface to 1,210 feet depth, gives the thickest and most complete section of the Dunvegan in the map-area.

The lower, or *Unio (P.)* cf. *dowlingi*, beds, interval 1,120 to 1,210 feet, consist of dark-grey shales, with much siltstones and less sandstones; few shell fragments are present in cuttings. The proportion of argillaceous cuttings increases downwards through this interval, marking the transition to the shales of the Fort St. John Group.

The remainder of the Dunvegan Formation contains brown, red, and olive coloured shales, grey and dark-grey shales, silty shales, carbonaceous shales, and shales with plant and coaly debris; siltstones; sandstones; and at several horizons thin coals, and thin grits or conglomerates. In the upper part, surface to 740 feet, shales, siltstones, and sandstones are in interbeds, and occur with several thin coal seams. In the lower part, to 1,120 feet depth, similar interbedded sequences are separated by sandstones in well-defined beds 30 to 40 feet thick, at intervals of 740 to 770, 790 to 820, 945 to 980, and 1,080 to 1,120 feet.

CHAPTER IV.—STRUCTURAL GEOLOGY

INTRODUCTION

The Rocky Mountain Foothills belong to the paratectonic system of the Western Cordillera. They were produced by the deformation and uplift of the sediments of a miogeosyncline or its extension. The complexity and closeness of folding and the displacements of faults decrease progressively northeastward from the Rocky Mountains through the Foothills to the Interior Plains. This decrease of deformation provides a basis for subdividing the Foothills belt.

The Rocky Mountain Foothills of the Peace, Moberly, and Pine River areas can be considered in two parts, from west to east: (1) the Inner Foothills, a belt of strongly folded, faulted, and disturbed rocks of Triassic, Jurassic, and Cretaceous ages; (2) the Outer Foothills, a belt of discontinuous folds and faults in which Cretaceous rocks form the main exposures. Deformation in the Outer Foothills was restricted to long faulted anticlines. A similar pattern of structures extends into the adjoining Plains, but the deformation was less intense. The divisions agree only partly with those proposed by Spieker (1920, p. 21). The Inner Foothills correspond to his " disturbed belt," and the Outer Foothills to his " moderately folded belt." Spieker's classification resembled that of Gwillim (1919).

The boundary of the Foothills and Plains is drawn along the east front of the Chetwynd anticline. It lacks any obvious lineament on the ground, but is appreciable from the different landscapes in views east and west from Mount Wabi. The common boundary of the Outer and Inner Foothills follows the line of the Pine River anticline along the escarpment made by the lower formations of the Fort St. John Group. The west boundary of the Foothills extends along the thrust at the east front of Solitude Mountain. West of this line Palæozoic rocks form the major outcrop in the Rocky Mountains.

Structures trend south 18 to 22 degrees east in the Outer Foothills and south 28 to 42 degrees east in the Inner Foothills. Older rocks are exposed in order from east to west. Folding is asymmetric; the axial planes of most large folds and the major thrusts dip southwest. In conventional terms, the rocks of the Foothills are overthrust from southwest to northeast. Thrust faults occur in both east and west limbs of the main folds. For purposes of description and following the convention, southwest-dipping thrusts and reverse faults are termed thrusts or faults of the " main sense." Correspondingly, northeast-dipping thrusts are designated " counter faults." Such faults are less common. They are of minor scale and restricted to the east limbs of the anticlines.

The following text describes structures in the Pine Valley. The Commotion structure is treated separately and in detail. Maps, Figure 2, and sections, Figure 3, illustrate the structural geology. Structures are very long and extend outside the map-area. For this reason, they are related to their extensions northwestward in the Foothills of the Moberly and Peace River areas in a following account of the tectonic framework; their extensions in the more remote area to the southeast are only known from reconnaissance (Muller, 1961). Folds in the map-area are of parallel type. They are of much interest, and require new descriptive terms and reclassification. The parallel folds include concentric, angular, cuspate, and lambdate forms. Appendix 1 treats the development of these fold forms with reference to examples in the Pine River area, and to others in the Moberly and Peace River areas where necessary. The patterns of deformation of the two divisions of the Foothills are different, and can be considered in further detail.

THE OUTER FOOTHILLS

Deformation and the main folding in the Outer Foothills were localized along the anticlines. The fold pattern of anticline and syncline is dissimilar, that is, discordant and incongruent.

The Outer Foothills have broad, shallow, open synclines of simple form and low amplitude. The synclines are 5 to 8 miles wide and alternate with narrow anticlinal belts which are about one-quarter of their width. The strata in the synclines are relatively undeformed, and they are depressed in order from west to east across the Foothills. The fold amplitude of the anticlines is much larger than for the synclines, and strata in anticlines are more strongly deformed and faulted. The anticlines may be simple or compound. They have concentric forms of low fold amplitude in the Pine River area; others in the Moberly and Peace River areas have angular or high amplitude concentric forms, or are modified by large thrusts. The east flanks of anticlines are broken by thrusts of the main sense directed against the adjoining synclines on the east. Anticlinal folding is asymmetric, and axial planes dip west.

The name Peace River structures is given for this regional configuration of anticlines and synclines. Tectonic movement involved compression in anticlinal folding and vertical displacement of the synclinal units, an uplift relative to the Plains. Displacement between synclinal units took place across the anticlines and their faults.

The Peace River structures of the Outer Foothills are discussed further in Appendix 2. Evidence only allows a limited appreciation of their structure at depth, and their origin is uncertain. Field work, drilled sections, comparative structures of foreland and paratectonic regions, and the tectonic framework of the Foothills indicate that the Peace River structures originated by block faulting and displacement in the basement during orogeny, rather than by dislocation of the sedimentary cover itself.

The Outer Foothills, where transected by the Pine Valley, are 25 miles wide and contain four major units of structure: the Chetwynd anticline, the Bissett syncline, the Commotion anticline, and the Hulcross syncline. The Wabi block borders the Chetwynd anticline on the east limit of the map-area. The Dunvegan and formations of the Fort St. John Group outcrop in the structures of the Outer Foothills. The Pine River anticline dividing the Inner and Outer Foothills has outcrops of the Crassier Group.

WARTENBE FAULT

The Wartenbe fault, lying outside the eastern boundary ascribed to the Outer Foothills, crosses the Pine River 5 miles northeast of Twidwell Bend. The fault trends about southeast. The northeast side, toward Mount Wartenbe, is downthrown with a stratigraphic displacement of the order of 1,500 feet, an estimate based on the exposures of the Dunvegan outcrops in the entrenched valley of the Pine. Williams (1939), Stelck (1941), and Stott (1961) noted this fault and the outcrop of post-Dunvegan Cretaceous strata, the Kaskapau Formation, at Mount Wartenbe. On the north side of the Pine Valley, a drift cover of silts, sands, and gravels obscures the trace of the fault.

WABI BLOCK

The Dunvegan Formation outcrops at plateau levels from the Wartenbe fault west to Chetwynd; the Cruiser shales outcrop in the valley. Exposures appear as a set of block structures, with strata about flat lying or tilted downwards to faults. The faults are of minor order, except for the Wabi fault, which trends south

85 degrees east across Mount Wabi. The fault itself is concealed in a draw, but alongside on the north, strata are downwarped against the fault. The block north of the fault is downthrown; the base of the Dunvegan Formation lies below 2,000 feet elevation, the level of the valley floor in Centurion Creek, and at 1,460 feet elevation in Sun et al Chetwynd 14-20; the strata are about flat lying. In the Wabi block, on the south side of the fault, the Dunvegan base attains elevations of 2,400 to 2,750 feet. The Wabi block is tilted with dips of 3 to 10 degrees southwest. The Wabi fault seems unusual because it is oblique to the structural trends of the Foothills and is not specially related to anticlinal folding.

CHETWYND ANTICLINE

The Chetwynd anticline trends south-southeast. Dunvegan beds formerly occupying the crest of the anticline are now breached and eroded. In projections the base of the formation attains an altitude of 3,050 feet at the crest along Windrem Creek, and about 1,350 feet higher than at the axis of Bissett syncline to the east. The Cruiser shales can be seen about Windrem Creek near its debouchement on the valley plain of the Pine. The Chetwynd anticline is concentric in form, with low fold amplitude; it is asymmetric in section with the axial plane dipping west-southwest. On the east side of the anticline the Dunvegan Formation is downthrown by 1,000 feet or more, measured across 500 feet horizontal distance, perpendicular to the fold axis. A thrust fault cuts the east limb of the anticline, or is so inferred. The projected trend of the anticline follows the Pine River along its course, downstream from Chetwynd to Twidwell Bend. Glacial-lake sediments, clays, silts, sands, and boulder clay, fill the valley in most of this reach. The few outcrops of Cruiser shales do not provide any clear record of the crest of the anticline or the fault on its east limb. A subordinate fault cuts the west limb of the Chetwynd anticline. The stratigraphic displacement of the Dunvegan sandstones is estimated at 400 feet. The fault is best shown as a thrust against the anticline, and this relationship is expressed in Figures 2 and 3. The Anglo Bralsaman Little Prairie No. 1 well entered its footwall.

BISSETT SYNCLINE

The Bissett syncline lies between the Chetwynd and Commotion anticlines, and is a broad structure trending south-southeast along its full length. Its east limb is flexured and has slight dips, less than 5 degrees. The west limb is steeper, and is broken by a minor fault with downthrow to the east, toward the synclinal axis about Bissett Creek. On the slopes of the Pine Valley, thick beds of Dunvegan sandstones form benches up to the plateau level at 3,000 to 3,500 feet. The base of the formation descends below 2,000 feet elevation about the synclinal axis. The syncline includes slight flexures marked in dips to 5 degrees. These are common in the Peace River structures and indicate minor warping or contraction within the synclines.

COMMOTION ANTICLINE

The Commotion anticline trends south-southeast. A fault system of parallel trend cuts its east front. Here, the Goodrich sandstones are turned down and thrust against Cruiser shales at Submarine Mountain and Young Creek. The escarpments of the Dunvegan Formation at Cruiser Mountain and east of Young Creek belong to the footwall part. The broadly arched form of the anticline is expressed by the massive conglomerate of the upper part of Member (iii) of the Commotion Formation. The conglomerate is exposed along the Hart Highway (Plate II) and along Commotion Creek upstream to the waterfall at 2,300 feet elevation. Above the waterfall, cutbanks reveal coal measures of Member (iv) of the Commotion Formation, and its contact with marine Hasler shales at 2,475 feet elevation. The Goodrich sandstones outcrop high in the valley sides at Commotion and Goodrich Creeks, and from here descend to the axis of the Hulcross syncline on the west. In projections, the Fort St. John-Dunvegan boundary, now eroded, reached elevations about 4,800 feet at the anticlinal crest (Fig. 3, Section NO, *see* the Commotion structure, following).

HULCROSS SYNCLINE

The Hulcross syncline trends south-southeast parallel with the Commotion anticline. The Pine River anticline, bounding the syncline on the west, swings to a more easterly trend. It constricts the width of the syncline from 9 miles about the north side of the Pine Valley to 4 miles on the south near Hasler Creek. At the axis of the Hulcross syncline, the Goodrich Formation descends to 2,300 feet elevation, 1,200 feet lower than the crest of the Commotion anticline, and the Fort St. John-Dunvegan boundary to about 3,750 feet elevation. Slight, almost indistinct flexures in the broad axial part of the syncline are recorded in Figure 2. They appear oblique to the main trend of the syncline, as determined from more extensive mapping. Minor flexures in synclines of the Outer Foothills are not easily defined for reasons of exposure, dependence on cross-bedded sandstones for observing slight dips, and the need for arbitrary placing formational boundaries at transitions.

PINE RIVER ANTICLINE (PART)

The Fort St. John beds rise in escarpments and cuestas along Crassier, Fred Nelson, Willow, and Browns Creeks, in the common limb of the Hulcross syncline and the Pine River anticline. The dips increase to 30 degrees in the westerly escarpments. The Pine River anticline is a compound structure. Its parts show fold forms and patterns characteristic of the Inner Foothills. No simple or unique structural break separates the Inner and Outer Foothills here (*see* Inner Foothills).

THE INNER FOOTHILLS

The Inner Foothills lie west of the Fisher and Crassier anticline in the Pine Valley and west of Carbon Creek farther north in the valley of the Peace River. They contain a series of closely folded anticlines, reverse faults, and few large thrusts. The anticlines and faults are closely spaced, and the widths of the synclines are much less than in the Outer Foothills. Wave length and amplitude of anticlinal and synclinal folding are about equal. Some of the synclines are modified by small-scale folding in the limbs. Others are narrowed and squeezed.

The sections of Figure 3 illustrate the structure of the Inner Foothills. Some regularity of the pattern of folding is apparent, and can be described in terms of a unit pattern with the following components: (1) The thrust, or thrust faulted anticline; (2) the closely folded syncline; (3) the rear anticline on the west, forming the footwall of the succeeding thrust or fault. This pattern is more obvious in the Rocky Mountains. It is less developed in the Inner Foothills, but it appears in the structures of Mount Bickford, Pyramis Peak, and Big Boulder Creek. Variations occur in the closeness of the folding and in the development of component folds and faults.

A distinction is made between the thrust faults of the anticlines and major thrusts. Thrust faults of the anticlines developed in folding; displacements are moderate and no extensive fault movements took place after folding was completed. The major thrusts underwent large movements, with stratigraphic displacement of 3,000 feet or more.

The fold patterns of the Rocky Mountains and the Inner Foothills are alike. The anticlines are more closely spaced, and the major thrusts are more important in the structures to the west. The Inner Foothills and the Rocky Mountains together can be regarded as one structural unit. In a structural sense, the placing of the boundary between them may be considered arbitrary.

The Inner Foothills show greater over-all and continuous crustal shortening than the Outer Foothills. There are no unfolded parts, and steeply dipping strata predominate.

PINE RIVER ANTICLINE (THE CRASSIER AND FISHER ANTICLINES)

Spieker (1920) first named the Pine River anticline. In later descriptions (Wickenden and Shaw, 1943; Spivak, 1944; McKechnie, 1956), the term refers to a compound anticline (or anticlinorium) which is deformed and folded in small and large-scale structures. The Crassier anticline on the east and the Fisher anticline on the west are its significant and major components. The Pine Valley contains the southern termination of the Fisher anticline and the northern termination of the Crassier anticline. The two anticlines overlap about their extremities. In this interval, in the Pine Valley, much subordinate folding and faulting complicates the structure. Some of this deformation was oblique and transverse to the main fold directions. Coal measures of the Crassier Group form the major outcrops in the Crassier and Fisher anticlines (Figs. 2 and 3).

The Crassier anticline extends about 2 miles north of the Pine River. On the river's north bank, part of the Dresser Formation is exposed in a concentric fold at the anticlinal crest, and again, farther northwest, overlying beds are folded in an angular form apparently about the same axis. The east limb of the Crassier anticline (common with the Hulcross syncline) is little complicated, while the west limb is faulted and modified by small folds. In the valley of Willow Creek, a lineament, marked by a string of narrow swamps, can be shown as a fault trace. In such interpretation it is a major fault, cutting obliquely across the anticlinal axis, and is of late development, perhaps post-dating the folding. On the southeast the outline of the Crassier anticline narrows, as its axis swings from south 35 degrees east at the Pine River to south 60 degrees east at the interfluve of Johnsen and Hasler Creeks. A few miles southeast of the Pine River, one major anticline is distinguished. A local, partial closure can be interpreted for the Crassier anticline about its intersection with the Pine River.

The Fisher anticline is angular. Its west limb dips uniformly at 58 degrees; dips decrease to 40 degrees near the axis; crest and axis coincide, and can be crossed in one pace. The east limb is less steep, and is modified by minor folds and faults. The axial plane inclines northeast, an unusual feature in the map-area. About 3 to 4 miles north of the Pine River, the Fisher anticline attains a culmination in which the Monach sandstones, Member (i), outcrop at 4,500 feet elevation. Seemingly a fault displaces the Monach quartzites, Member (ii), on the east limb of the anticline, close to the axial plane. The Fisher anticline plunges steeply southeast. Down the plunge, on the north side of the Pine Valley, its structure becomes complex, and greatly altered by folds and faults developed in the Crassier coal measures. These continue across the valley, but farther southeast the Fisher anticline loses its identity. Structures of the Crassier and Fisher anticline merge, as the compound Pine River anticline narrows in width and alters in trend. The northwest plunge of the Fisher anticline is slight, but its extent is unmapped. The anticline is continuous with the Frank Roy anticline, or part of this compound structure, to the northwest across the Moberly River.

FISHER SYNCLINE

The Fisher syncline, separating the Pine River and Bickford anticlines, trends south 30 degrees east along Fisher Creek, swinging to south 55 degrees east along

Falls Creek on the southeast. The syncline plunges southeast, the Gething-Moosebar boundary along the axis declining 1,000 feet in the mapped area. The Moosebar outcrop widens to the southeast, corresponding to the plunge of the syncline and of the Fisher anticline adjacent to Falls Mountain. Sandstones of Member (i) of the Commotion Formation make an outlier at Falls Mountain. The Fisher syncline contains many subsidiary folds, faults, and shear zones. Thrust faults repeat the Gething-Moosebar contact almost everywhere along its outcrop, with stratigraphic displacements ranging from 1 to 600 feet (Fig. 2). Plate III shows an example of this prevalent faulting.

BICKFORD ANTICLINE

The Bickford anticline is a large angular fold, split by a fault about the axial plane. Its west limb is overthrust. The Monteith, Beattie Peaks, and Monach Formations are brought up in steeply inclined limbs, the west limb at dips of 40 to 70 degrees, and the east at 60 degrees to about vertical. The form of the anticline is less acute northwest and southeast from the summit ridge of Mount Bickford. The great amplitude of the fold and the relief of the Bickford ridge give the impression of a buttress. Along the ridge, the axial fault is shown by minor angular dragfolding in the Monteith quartzites, and by displacement of the projected Monteith-Beattie Peaks boundary in sections across the axis. A corresponding axial fault is visible in the south wall of the Pine Valley. Here, in the west limb of the anticline, a sector of dislocation and rumpling may involve repetition of Monteith beds, but no clear proof was obtained. This dislocation does not appear farther northwest. Shales in isolated exposure about the anticlinal axis, on the north wall of the valley near the river plain, may belong to the Fernie Group. The Bickford joins with the Monteith anticline on the northwards, to form a long compound anticline extending to the divide between Carbon Creek and the Moberly River. One of the component anticlines borders the Bickford anticline on the east, and contains Beattie Peaks and Monach outcrops in its core, at 5,000 feet elevation near Bickford Lake. This component anticline plunges southeast, and loses definition in minor folds and faults developed in the Crassier coal measures about the Pine Valley.

COYOTE CREEK SYNCLINE

The Coyote Creek syncline, a narrow compressed structure between the Bickford and Big Boulder anticlines, contains subsidiary folds of parallel trend in the Crassier coal measures. These folds have angular and accordion forms. They are thought to overlie a concealed detachment plane that separates them from underlying beds of the Crassier Group, which are in continuous, unfaulted, and parallel order with Beaudette strata in the west limb of the Bickford anticline.

BIG BOULDER ANTICLINE

The Big Boulder anticline is a concentric fold of high amplitude. Its box form is shown by the core of Monteith strata and the wide arch of the Beattie Peaks and Monach outcrops. On the north wall of the Pine Valley, the east limb of the anticline is attenuated, and incomplete outcrops of the Beattie Peaks and Monach Formations indicate a thrust fault, downthrown on the northeast. The fault is inferred on the southeast, where the Coyote Creek syncline is more open, and where the Beattie Peaks and Monach Formations and lower beds of the Crassier Group are complete and present at the level of the river plain. The Big Boulder anticline plunges northwest; the Crassier Group occupies its crest, away from the Pine Valley. The west limb of the anticline dips at 40 to 65 degrees from a sharp inflection or fold break with the broad crest. Southwest of the anticline, the Crassier coal measures are deformed in close folds, which represent parts of a compound syncline lying near the footwall of the Pyramis thrust. Some folds are angular, and others concentric or combined concentric-angular (cuspate). The Monach sandstones and quartzites are brought up to the level of the river plain in the cores of few anticlines. A narrow segment of Beaudette strata forms the footwall of the Pyramis thrust.

PYRAMIS THRUST AND SYNCLINE

The Pyramis thrust cuts shales in the upper part of the Fernie Group. The shales and overlying sandstones of the Beaudette Group comprise the overthrust block, and dip about 54 degrees southwest, parallel with the thrust. Stratigraphic displacement at the thrust amounts to 4,000 feet, and movement exceeds 7,500 feet (dip slip). Fernie and Beaudette strata are folded in the Pyramis syncline west of the thrust. Subordinate folds and faults, mapped and inferred in the east limb of the syncline, compose an intricate pattern of dislocations in the overthrust. On the west, in the adjoining Silver Sands anticline, Fernie and Beaudette overlie Triassic strata in normal order without fault or thrust separation. About 4 miles northwest of the Pine River, near the map limit, are exposures of thinly interbedded shales, siltstones, and sandstones, with one layer of cobbles, which are referred to the Crassier Group. They are dragfolded in the west limb of the Pyramis syncline, but other aspects of their field relationships are not revealed here.

CAIRNS AND SILVER SANDS ANTICLINES: LE MORAY SYNCLINE

North of the Pine River, the Cairns and Silver Sands anticlines and their intervening structures are close folded and faulted. Calcareous, dolomitic siltstones and sandstones, and dolomites of lower parts of the Grey Beds are found in the anticlinal cores, as in cuts along the Hart Highway. Limestones in the upper part of the Grey Beds outcrop on the north wall of Pine Valley, and also at the terminal face of the long ridge, extending from the north salient of Mount Le Moray to Mountain Creek, near its confluence with the Pine River. The ridge contains an angular fold, illustrated in the section, Figure 21. The Pardonet Formation and its sequence of faunas can be traced in the west limb of Silver Sands anticline and partly at Cairns Creek. The *Monotis* zone is a useful guide in mapping structures on account of its frequent exposure (*see* Pardonet Formation). The Fernie Group mostly occupies synclines or the footwalls of thrust faults. Its few exposures show small-scale folding and faulting, and some cleavage in places.

Le Moray syncline has a simple open form in the competent Beaudette sandstones. The syncline overlies a narrow, compressed, probably faulted syncline formed in Triassic beds. Fernie shales separate these disharmonic folds, and are deformed in thrust faults and shear zones about the fold axes. The upper syncline plunges southeast at the north salient of Mount Le Moray. It has no obvious counterpart in structures on the north side of the Pine Valley where Beaudette strata are absent, even at higher levels (4,000 feet elevation). Structures of Triassic outcrops can be correlated from northwest to southeast across the Pine Valley, assuming some divergent fold trends, and excluding one or two folds. The structures indicate a general plunge or depression to the southeast.

SOLITUDE THRUST

Mississippian shales and limestones outcrop at the base of Mount Solitude. They belong to an overthrust which includes Upper Palæozoic and Triassic strata. The Solitude thrust serves a conventional boundary-line, dividing the Rocky Mountains from the Foothills on the east. Beds in the overthrust dip 50 to 55 degrees southwest. Southwest from the thrust they form a syncline and anticline, next to another major thrust along Callazon Creek. The Solitude thrust has stratigraphic displacements of the order of 4,000 to 6,000 feet, and corresponding movement more than 8,000 to 10,000 feet. Structures in its footwall make a complex of tight folds and thrusts, by which Grey Beds are brought above the Pardonet, and these, in turn, above Fernie strata. Road and railway cuts at the West Pine Bridge reveal some details of the structures (*see* under Pardonet Formation).

THE COMMOTION STRUCTURE

Introduction

The Commotion structure was first described by Spieker from a reconnaissance of the Foothills south of the Peace River. In his report of 1920, Spieker named the structure the " Boulder Creek Dome " (page 22). He noted a seepage of highly inflammable sulphurous gas issuing from sandstones underlying the " Boulder Creek " conglomerate on the west flank of the structure.[*] Spieker considered the structure favourable for drilling.

The area was prospected by M. Y. Williams in 1938 and 1939, who reported a dome-like appearance for the structure at surface and a fault on the east side. Following the explorations, the Government of British Columbia drilled the Pine River No. 1 well on this anticline, near the confluence of Commotion Creek and the Pine River. The drilling ran into coal measures of the Crassier Group at 2,424 feet and kept in the same beds to final depth at 6,941 feet. Cores taken from the middle section of the well showed steep and vertical dips. The drilling was suspended in 1942, without encountering oil staining or shows. No formation tests were run.

During the drilling of the well, some doubts had occurred about the true nature of the structure, and it became clear that the age of the beds on the east and west sides of the anticline was open to question. The stratigraphy was established by Wickenden and Shaw (1943). Their mapping showed the Dunvegan Formation on the east side of the anticline, and that the eastern part, comprising the Dunvegan Formation and the Cruiser Formation of the Fort St. John Group, was in faulted contact with the Goodrich Formation on the west. In a brief discussion of the structure, Wickenden and Shaw concluded that a plane of rupture occurred on the east flank of the fold, but that it was " not apparent from the displacements of the beds on either side that the fault represents a thrust from the west." They suggested that the ruptured beds lay along a narrow compressed syncline parallel with the anticlinal axis, and lying between Submarine and Cruiser Mountains.

The Commotion structure was reviewed briefly by Hume (1941, p. 35) before it was tested, and later discussed by Stelck (1941, p. 80), and by McLearn and Kindle (1950, p. 118).

Remapping of the structure was proposed in the work on the Pine River Foothills, and carried out in 1955 and 1956. The structure has been examined by combining surface mapping and data available from the drill cuttings, cores, and the well records. The subsurface structure is not completely proved by one well, but can be indicated within limits of available data. The Commotion structure is deceptive. The surface and subsurface forms are unlike.

General Description

The Commotion structure is shown in Figures 2 and 3. It lies in the Outer Foothills of the Pine Valley. The anticline is of low amplitude. Folding at the surface is concentric, and the beds are broken by faulting on the east limb. The

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[*] The " Boulder Creek " conglomerate and underlying sandstones are known now as the upper marine division of the Commotion Formation, Member (iii). Spieker did not specify the locality of the gas seep any further, and it has not been mentioned by subsequent authors. No gas seeps are now known in the Commotion Creek area.

Commotion anticline can be recognized for a length of 12 miles. It separates two major flat synclines—the Hulcross syncline on the west and the Bissett syncline on the east. The east syncline is depressed some 2,000 feet relative to the Hulcross syncline. The rocks exposed at the surface include the Dunvegan Formation, and the Cruiser, Goodrich, Hasler, and Commotion Formations of the Fort St. John Group. The underlying beds, the Moosebar Formation of the Fort St. John Group, and the Crassier Group, were drilled in the Pine River No. 1 well.

The oldest exposed beds are conglomerates and sandstones of the upper part of the Commotion Formation, Member (iii). They are in continuous exposures in the lower sides of the Pine Valley, forming a broad, low, flat-topped arch over a span of 2 miles. The arch is carried through similarly in the trace of the Goodrich sandstones in the higher shoulders of the valley. The structure on first appearance resembles a gentle anticlinal warp, with some slight asymmetry of the east limb.

The traces of the beds, observable in views from the north and south sides of the Pine Valley, are shown in Figure 3, Section NO. The west limb of the anticline rises at a low angle (dip, 2 degrees), and is marked off from the Hulcross syncline by a slight inflection of the beds and an increase of dip to 8 degrees. The middle part is a wide platform at the crest, with dips sensibly horizontal for distances of 3,000 to 6,000 feet on the north and south sides of the valley. The anticline terminates sharply on its east limb. A brief change of dips occurs, and they increase eastwards to the order of 40 to 60 degrees, where the upper beds of the Commotion Formation and the Hasler and Goodrich Formations can be traversed within a distance of 2,500 feet on the approach to Young Creek at its debouchement on the plain of the Pine Valley. The east flank of the anticline is faulted, and the fault is exposed in Young Creek. The sharp down turning of the beds can be also seen on Submarine Mountain facing the Pine Valley. Eastward from the descent of these beds, the Goodrich and the Cruiser Formations occur in poor exposure in the valley sides. They are overlain by the Dunvegan Formation forming a high table-land sloping to the east. These beds on the east side of the structure form the west limb of the Bissett syncline.

Mapping indicates subordinate anticlinal flexures along the crest of the Commotion structure. They are marked separately by axes in Figure 2. The trace of the Hasler-Goodrich boundary allows possible closures in the order of 100 to 200 feet, but the individual closures are not defined in full. Similarly, a culmination of the entire Commotion anticline remains unproven in the map-area.

The surface structure illustrates the relationship of the Commotion anticline and the Bissett syncline. These two parts are not in continuous simple structural order. There is a gap between them, marked by the col between Submarine and Cruiser Mountains, and by Young Creek on the south side of the Pine Valley. The beds of the two parts do not match across the gap. They are separated by a structural break.

RELATIONSHIP BETWEEN THE COMMOTION ANTICLINE AND THE BISSETT SYNCLINE

The relations between the main Commotion anticline and the Bissett syncline are most clearly shown at Young Creek, on the south side of the Pine Valley. The structural break between them is shown on Figure 6.

In the lower reach of Young Creek the Commotion anticline is overthrust against a complex of dragfolds developed in the Cruiser Formation. This complex is in turn overthrust on its east limb. It then forms a compressed core which is overridden by the thrust fault of the Commotion anticline in the main sense, and the counter thrust of the west limb of the Bissett syncline. In the west fault the Goodrich Formation is thrust against the Cruiser beds. The fault is visible near the confluence of the two main tributaries of Young Creek. On the east side across Young Creek to Caron Creek, the Cruiser beds occur in repeated order.

A corresponding structure between the main Commotion anticline and the Bissett syncline is found on the north side of the Pine Valley. It can be partly seen about the col between Submarine and Cruiser Mountains, in traverse along the first tributary on the east side of Commotion Creek. The Cruiser shales are exposed in the upper course of the tributary stream, where they are broken by small faulting, rumpling, and dragfolding (Fig. 3). The small folds have axial planes dipping 70 degrees southwest. The disturbed beds of the Cruiser Formation may be in structural continuity with the Dunvegan outcrop of the Bissett syncline on Cruiser Mountain. Otherwise, they may be separated by a fault of slight displacement. Elsewhere, there is no continuous exposure showing the relation of the Commotion anticline and the Bissett syncline, and the structural break between.

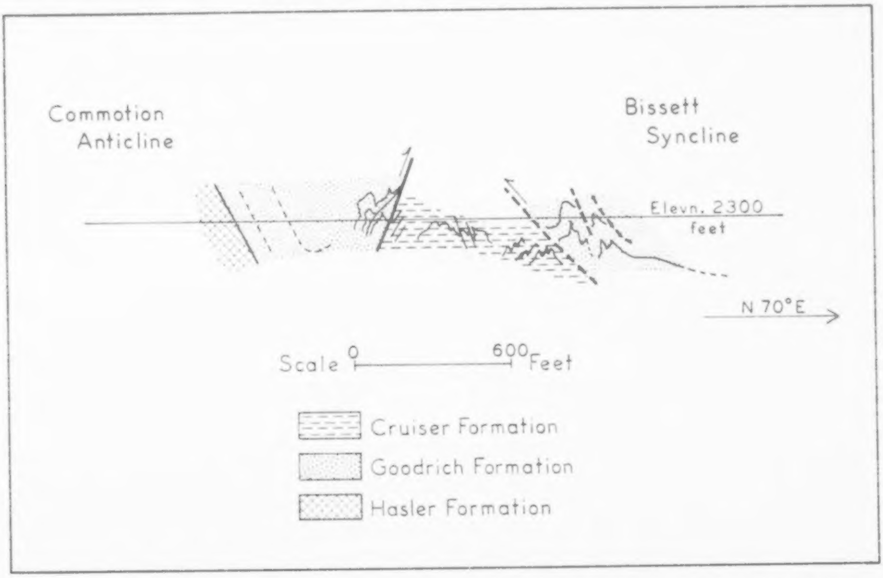

Figure 6. Diagram of the field relations of the Commotion anticline and Bissett syncline at Young Creek.

In summary, a structural break composed of thrust faulting and dragfolding separates the main Commotion anticline from the Bissett syncline on both sides of the Pine Valley. The stratigraphic displacements across the break are small. Estimates of less than 250 feet were obtained in the mapping. The relative depression of the Bissett syncline is provided by the dip of the beds in the west limb of the Commotion anticline and the east limb of the Bissett syncline (Fig. 3).

The Northward Continuation of the Commotion Structure

In its northern extension the Commotion structure ends against the plunging termination of the Hulcross anticline.

Northward from the confluence of Walton and Commotion Creeks, the structure yields to a monoclinal fold according to the trace of the Goodrich Formation. A minor counterthrust mapped in the Dunvegan Formation is directed against the monocline (Fig. 2). The Bissett syncline is thrust westward on this fault. Most of the structure of the footwall is concealed. North of Submarine Mountain the detail of structure between the main Commotion anticline and the Bissett syncline is also concealed. Some faulting is inferred from the trace of bedding on air photographs taken over freshly burned ground. It seems that no unique major fault is necessary in mapping the northward continuation of the Commotion structure.

THE SOUTHWARD CONTINUATION OF THE COMMOTION STRUCTURE

The main Commotion anticline extends southward to the middle part of Caron Creek. The Bissett syncline continues south from the Pine Valley also. The distinction of the two units is maintained. They are separated by a series of structural dislocations in the high table-lands south of Young Creek and in the valley of Caron Creek. Few exposures occur across this ground.

The continuity of the structural break described in the lower part of Young Creek is lost in this area. A short fault, or a monoclinal flexure, was mapped south of the headwaters of Young Creek. This structure displaces the Goodrich Formation, and the east side is depressed. Faulting is also inferred in Caron Creek. It is indicated by a small inlier of the Hasler Formation, found in one exposure along the bank of the creek.

A partial closure on the southeast of the Commotion structure is shown by a lapping round of the Dunvegan outcrop and a definite reflection of the prevailing strike at Caron Creek. The reflected strike corresponds to the stagger of minor structures in echelon between the Pine Valley and Caron Creek. These minor structures may mark the loci of the major differential movement in the upper parts of the structure. They can be considered to delineate the front of the Commotion anticline and its boundary with the Bissett syncline. The front of the Commotion anticline is then recurved back to the west, in its southern extension. The continuity of the anticline is maintained without any apparent plunge to the south.

THE PINE RIVER NO. 1 WELL

Descriptions of drill cuttings and cores are given in Appendix 3. The deviation of the drill-hole, recorded in Appendix 3, is taken from a compilation graph of engineering data concerning the well by T. B. Williams, petroleum engineer in charge (1942). No electric logs nor any other kinds of hole survey were run.

Bedrock was entered at 1,081 feet. The Crassier-Fort St. John contact, marked by the basal conglomerate of the Moosebar Formation, is found at a depth of 2,424 feet in drill cuttings. This contact is the only precise datum horizon obtained in the well section. Thereafter beds of the Crassier Group were followed in the drilling for some 4,500 feet vertical distance. Stratigraphic units can be assigned to this interval as follows:—

Depth (Ft.)	Stratigraphic Units
2,420 to 5,410	Gething Formation.
5,410 to 5,730	Gething Formation — lowermost beds, and transition to Dresser Formation: or ?Dresser Formation.
5,730 to 6,941 (total depth)	Gething Formation.

The Dresser Formation is distinguished by its greater proportion of sandstones, and its interbeds of medium- to very coarse-grained sandstones, grits, and sporadic conglomerates. These features continue upwards in the transition with the Gething Formation. The Dresser-Gething boundary needs to be decided in complete sections or judged with some arbitrary measure. In Pine River No. 1 well, the 75 feet of coal measures with coarse-grained sandstones and pebbly layers in the interval 5,410 to 5,730 feet may seem insufficient to fix the Dresser-Gething boundary.

It is possible to discriminate the upper, middle, and lower parts of the Gething Formation from proportions of shales, of siltstones, and of sandstones of different grain sizes. The sand/shale ratio and the proportions of the coarser fractions in the sandstones tend to increase downwards in the Gething Formation. Such criteria are used to interpret the subsurface structure (Fig. 7), but they allow only general inferences to be made. From the drill cuttings it appears that coal measures of the upper half of the Crassier Group are repeated below a depth about 3,500 feet, and again repeated below a depth within the interval 5,750 to 6,250 feet. It is estimated that the drill cut 300 to 400 feet of strata in the vertical interval 4,600 to 6,200 feet, according to the dips shown in the cores. The drill cuttings support this deduction, and, in illustration, one coal seam was followed for 80 feet vertical distance. Deviations of the hole range from 1 to 10 degrees. Coal seams of the Crassier Group rarely exceed a thickness of 20 feet (inclusive of bone and non-coaly partings), and the maximum thickness recorded for groups of closely spaced seams is less than 30 feet (McKechnie, 1955, p. 29).

The cores taken in the Crassier beds consist of an upper group of low to moderate dips; a middle group of steep and vertical dips, 65 to 90 degrees; and a third group of moderate dips (Appendix 3). The distribution of the dips is summarized as follows:—

Depth (Ft.)	Dips in Cores
2,915 to 3,909	3 to about 30 degrees.
4,662 to 5,907	65 to 90 degrees; some indeterminate dips.
5,907 to 6,249	Cores missing.
6,249 to 6,853	5 to 30 degrees.

The cores are jointed and fractured. Many fracture planes bear slickensided and polished graphitic surfaces. Tension fractures occur and are infilled with quartz and carbonate veining. Veining is frequent, and vein material is abundant in drill cuttings for much of the section in the Crassier beds. The data provided by the cores are open to several interpretations. The cores are not oriented, and the bedding may dip east or west and may be in normal or inverted order.

REMARKS: RELATIONSHIP OF SURFACE AND SUBSURFACE STRUCTURE

The Commotion structure belongs to the Peace River structures of the Outer Foothills. Its characteristics may be reviewed.

(1) At the surface it is a simple anticline of low amplitude and broad span. The east limb turns down sharply, making an abrupt change of folding.

(2) There is a definite structural break separating the Commotion anticline from the Bissett syncline. Detail of this structural break is mostly concealed. It includes a group of small folds and thrust faults of the main and counter sense.

(3) The Pine River No. 1 well proves a long steep limb of a fold at depth; and below this, folding with moderate dips.

(4) The surface and subsurface folds are unlike and do not match.

(5) The Commotion anticline is asymmetric. The axial plane is indefinite at the surface due to the broad flat crest. Its general trace in the structure can be indicated. The axial plane flattens from the surface structure to depth.

(6) The east front of the Commotion anticline at the surface is displaced far eastward of the folding proved at depth in the Pine River No. 1 well. This displacement emphasizes the difference between surface and subsurface structure.

The development and mechanisms of folding of the Commotion structure are discussed in Appendix 1. The fold pattern is not unusual (*compare* the Tarra anticline, and the Sarrebrück anticline, de Sitter, 1956, p. 244). This type of folding occurs commonly with the following geological conditions: (1) Thick sequences of incompetent beds, shales, mudstones, etc., containing one or two competent formations; (2) a limiting of the compression—the compression does not become far advanced. Such conditions existed for the Commotion structure. The separation of the surface and subsurface structure took place in the shales of the Moosebar Formation. In the Outer Foothills, limited lateral contraction produced folding in the competent Commotion and Goodrich Formations, while lower in the sequence complex disharmonic structures seem to have developed in the coal measures of the Crassier Group.

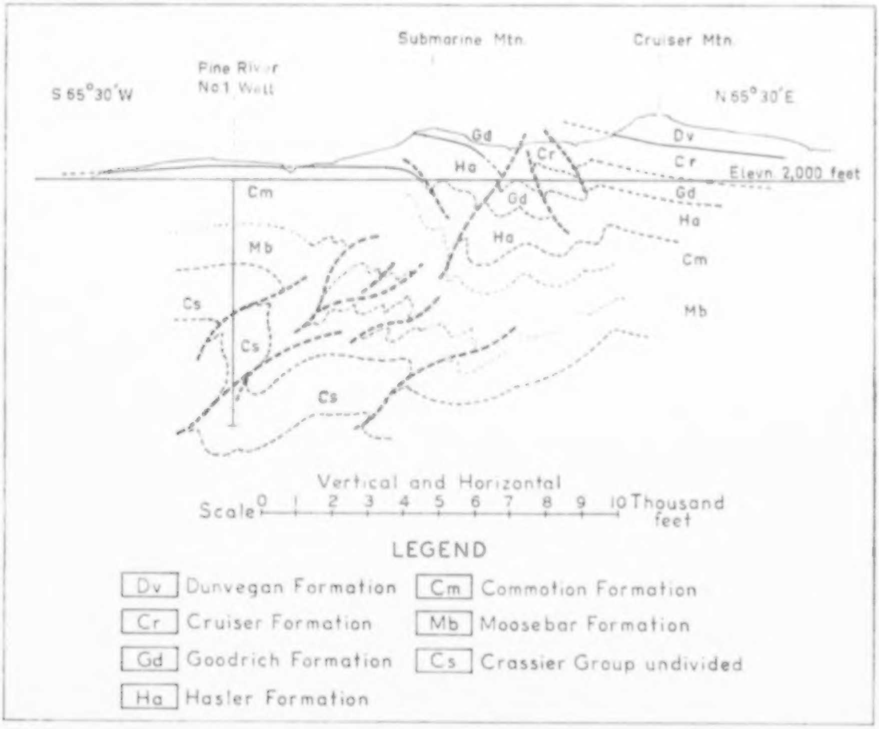
Submarine Mtn.
Cruiser Mtn.
Pine River No.1 Well
S 65°30'W
N 65°30'E
Dv
Cr
Gd
Ha
Cm
Mb
Cs
Elevn. 2,000 feet
Vertical and Horizontal
Scale 0 1 2 3 4 5 6 7 8 9 10 Thousand feet
LEGEND
Dv Dunvegan Formation
Cr Cruiser Formation
Gd Goodrich Formation
Ha Hasler Formation
Cm Commotion Formation
Mb Moosebar Formation
Cs Crassier Group undivided

Figure 7. Interpretation of the Commotion structure.

The Pine River No. 1 well allows the Commotion structure in the subsurface to be interpreted as follows (*see* Fig. 7). In the drilled section, the fold in the Crassier Group has a sigmoid form. The drill cut the vertical limb of a syncline; tops of beds in this limb are supposed to face eastward. The syncline is narrow and probably compressed against an anticline on its east side. Two thrust faults, each repeating parts of the Crassier Group, are believed to intersect the drilled section of the well. One fault, at a depth of 3,500 feet, overlies the vertical limb of the syncline. The lower fault, at about 6,000 feet, overlies the axial part of the syncline. Minor faults may also be present in the well section. Within the syncline, the Moosebar shales (Fort St. John Group) are cut by a series of thrust faults. A complex of minor folds is developed over the fault planes. The movements on the fault planes are not extensive, and they probably diminish and pass into folding in the shales. Folding above local minor décollements may also be present in the shales. In another view, a single major thrust fault may be substituted for this complex of thrusts and minor folds, and décollements. The remaining beds of the Fort St. John Group are folded and jammed in a minor anticline which is supposed to underlie the structural break between the Commotion anticline and the Bissett syncline. This minor anticline is thrust faulted on both east and west sides, according to the surface mapping at Young Creek.

The Commotion structure indicates the importance of seismic surveys in exploration preceding drilling. The form of the structure in the subsurface cannot be predicted from surface mapping, and geometric projections of the structure at depths are unreliable. The axial plane of the anticline at the surface may be discontinuous with the underlying folds, or its dip may decrease at depth. The surface anticline extends far over the underlying folding in the subsurface section of the well. This indicates the presence of décollement or low-angle thrusting in the Moosebar shales. Moreover, the east front of the anticline is not cut by one continuous simple thrust fault, as shown from the surface mapping. It is inferred that the fold compression was taken up by complex small folding and thrust faulting in the shales and mudstones—the structural break between the Commotion anticline and the Bissett syncline. Such inferrence supports Wickenden and Shaw's view that "it is not apparent from the displacement of beds on either side that the fault represents a thrust from the west."

It is possible to reconstruct aerial sections showing the Dunvegan Formation in a counter thrust on the east flank of the structure. The counter thrust can be drawn passing across and above the structural break between the Commotion anticline and the Bissett syncline. One counter thrust forms a component of the structural break at Young Creek; another is mapped in the Dunvegan Formation (Fig. 2). Such counter thrusting is directed against the axes of anticlines of this type, but probably only at some stratigraphic levels. Counter folds and thrusts may then appear deceptive, but they provide subsidiary structures for the most part.

FOLDING, FAULTING, AND TECTONICS

Folds in the map-area are of parallel type. The faulting was contemporary and related to the folding. Both result from compression and uplift in the Rocky Mountain orogeny, a phase of the Laramide revolution. The age of the orogeny is regarded as post-Cretaceous, though no direct dating is available in northeastern British Columbia. In Alberta, major uplift on the site of the Rocky Mountains occurred in Eocene-Early Oligocene time (Russell and Wickenden, 1933; McLearn and Kindle, 1950; Russell, 1951, 1954; Tozer, 1956).

Fold Forms

The parallel fold forms are concentric or non-concentric. Concentric folds are of low or high fold amplitude; non-concentric (angular) folds include those of cuspate and lambdate forms. The classes of fold forms are determined from right sections, perpendicular to the fold axis, for the bedding planes selected. This classification derives from the mapping, and is stated more completely in Appendix 1 (*see also* Fig. 8).

Concentric folds have strata bent in arcuate form, about a common fold centre, in right section. Concentric folds of low amplitude have low dips, less than 12 degrees, for strata at the point of contraflexure between syncline and anticline. In concentric folds of high amplitude, the corresponding dips exceed 40 degrees. There are no intermediate forms, a relationship indicating different modes and environments of folding for each. Low amplitude concentric folds belong to the Outer Foothills, the Commotion and Chetwynd anticline being typical, and the only examples from the map-area. High amplitude concentric folds occur in the Inner Foothills, and the Big Boulder and Cairns anticlines are examples. Appendix 1 explains the development of these fold forms.

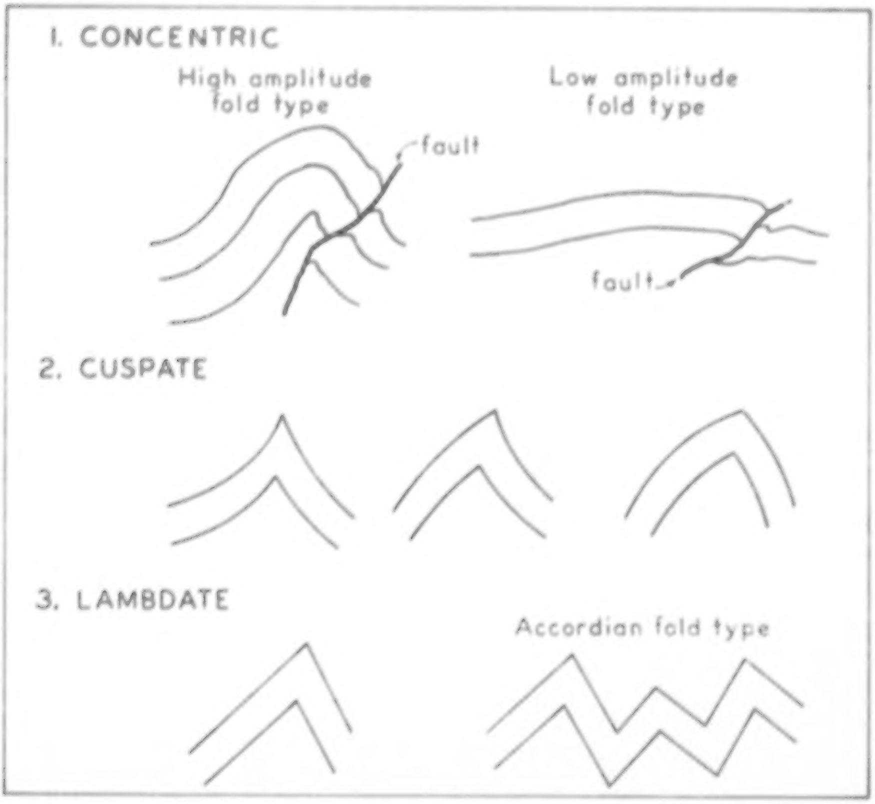

Figure 8. Classes of parallel fold forms, illustrated in right sections.

Cuspate folds have arcuate strata bent in a sharp angular fold across the axial plane. The Noman anticline, a subordinate fold flanking the Bickford anticline, is a distinct example (*see* Fig. 25, taken from serial sections in McKechnie, 1955). Several folds denoted as angular may be redescribed as cuspate, from more critical study, by opinion, or by resource to theory.

Lambdate folds are parallel folds having straight limbs, with a sharp angular fold across the axial plane. Lambdate forms commonly occur in a set of folds (accordion folds). Examples from the Pine Valley area are few, and all are subordinate folds adjacent or related to thrust faults or detachment planes, for example, minor folds in the Crassier coal measures in the Coyote Creek syncline, and folds of the Beattie Peaks Formation in the west flank of the Bickford anticline southeast of Bickford Lake.

Distinction of cuspate and lambdate folds may be uncertain or impracticable in the field, for lack of exposure and detailed measurement. The term angular describes folds where cuspate and lambdate forms are not so distinguished. The Bickford and Fisher anticlines, and the anticline in the Grey Beds near the confluence of Mountain Creek and the Pine River, are denoted as angular folds.

Some folds may be composite, with concentric and cuspate (or angular) forms at different levels and combined about the same general axis of folding. The northwest termination of the Crassier anticline north of the Pine River and structures west of the Big Boulder anticline provide likely examples. Most angular folds, the cuspate, high amplitude concentric, and composite fold forms occur together in the Inner Foothills. Their developments and relationships are treated in Appendix 1.

FAULTS AND THRUSTS

Faulting is prevalent as a release of strain and accommodation of volumes in the former folding compression. Faulting occurred within the folds and modified their forms. Thrust faults cut the steep eastern limbs of concentric anticlines. They share a parallel trend and override the synclinal limbs on the east. Most angular, cuspate, and lambdate folds have faults, about parallel and close to the axial plane; for these, the fault displacements are not large relative to the fold amplitude, and the sense of movement is not obvious in every fold. Also, thrust faults of minor order are contained in the limbs of concentric and non-concentric folds.

The major thrusts, the Pyramis and Solitude Mountain thrusts, differ from the thrust faults of anticlines by their greater stratigraphic displacement, of the order of 3,000 feet or more, and by their greater movement (dip slip) in excess of 5,000 feet. The thrusts follow the structural trend and cut the Jurassic and Carboniferous shales respectively. The overthrust blocks have simple structure, with thick competent strata dipping about 55 degrees southwest, parallel with the thrust planes. The overthrust blocks pass to folds to the southwest, in the rear of the thrusts. Structures in the footwalls of the thrusts are complex and not resolved in detail everywhere. They provide little evidence for assessing movement on the thrust or its sequence of development.

TECTONICS

Structures in the Pine Valley belong to a tectonic framework which includes the junction of the northern and southern parts of the Rocky Mountains and Foothills, and in the Interior Plains, the junction of three major units of the foreland, the Halfway block, the Fort St. John arch, and the Alberta syncline. This tectonic framework and its history was examined by the writer (MSS., 1963). The following note summarizes the conclusions from this work, and Figure 9 illustrates the tectonics.

The northern and southern parts of the Rocky Mountains and Foothills form a continuous fold system. They are distinguished by different fold trends, and a different alignment north and south of latitude 54 degrees to 55 degrees 30 minutes north.

The tectonic units of the foreland occupy the Interior Plains for a width of 100 miles from the boundary with the Foothills. They were differentiated in the Laramide revolution. Mapping on horizons in the Lower Cretaceous of the Fort St. John Group defines these units. The Halfway block corresponds to the northern Rocky Mountains and Foothills. It contains a series of longitudinal folds; these resemble the Peace River structures, but the deformation and displacements on faults are less. Cretaceous horizons show a northeast tilt for the Halfway block, opposite to the uniform homoclinal dip for the rest of the Interior Plains. The Alberta syncline is a structural depression, contained by an increase of the south-

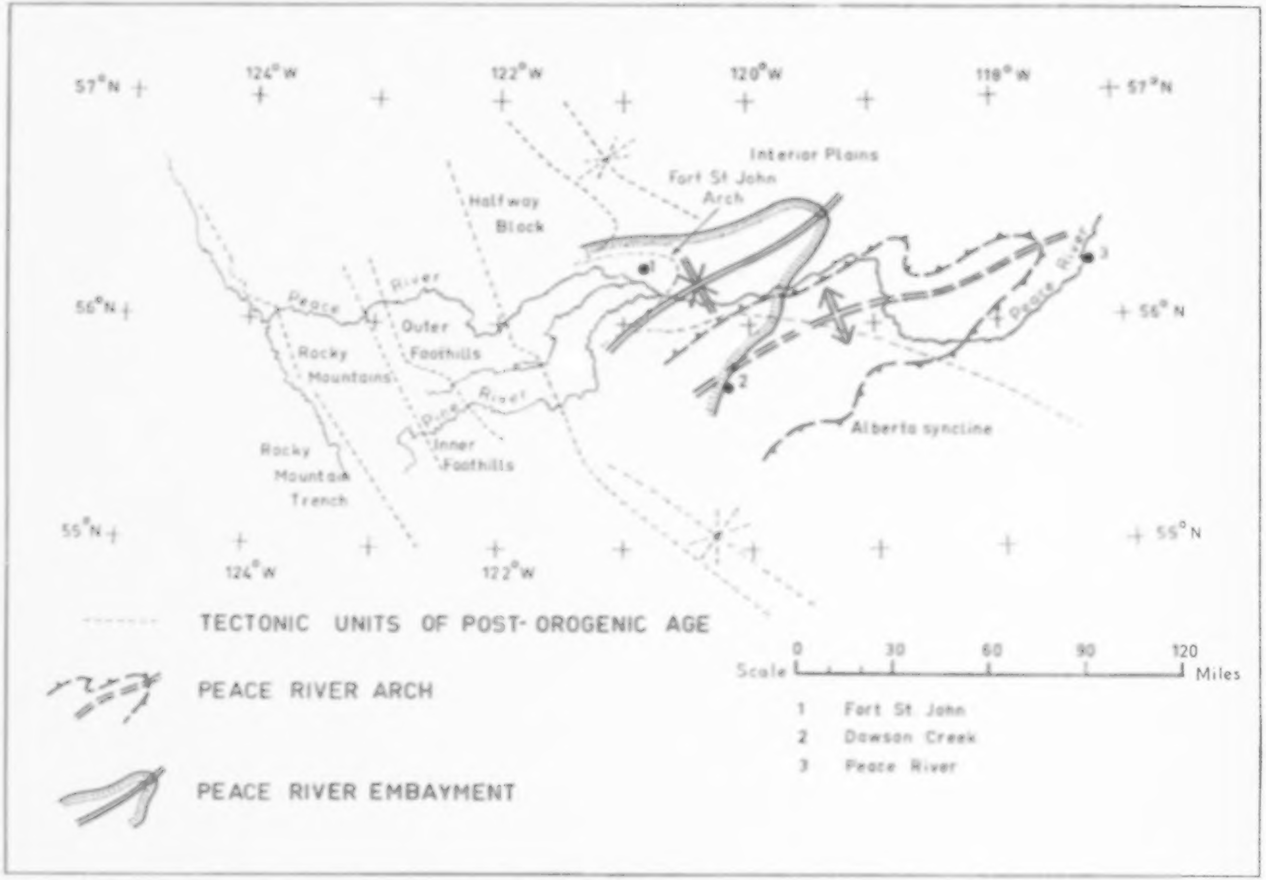
57°N
124°W
122°W
120°W
118°W
57°N
Interior Plains
Fort St John Arch
Halfway Block
56°N
56°N
Peace River
Outer Foothills
Rocky Mountains
Pine River
Inner Foothills
Rocky Mountain Trench
Alberta syncline
Peace River
55°N
55°N
124°W
122°W
TECTONIC UNITS OF POST- OROGENIC AGE
PEACE RIVER ARCH
PEACE RIVER EMBAYMENT
Scale
0
30
60
90
120
Miles
1 Fort St John
2 Dawson Creek
3 Peace River

Figure 9. Tectonic environment of the Pine Valley.

west monoclinal dip in the Plains and the uplift of the Foothills. The Alberta syncline represents the foreland of the southern Rocky Mountains and Foothills. The Fort St. John arch separates the Halfway block and the Alberta syncline. It extends from the Foothills to Fort St. John, about an axis trending north 65 degrees east; uplift on this trend persists eastward to longitude 120 degrees, at Boundary Lake.

The junction of the tectonic units of the Plains now occupies the site of the former Peace River embayment. This was a basin of differential subsidence and sedimentation from Mississippian to late Cretaceous times. The embayment received a great thickness of sediments. About 10,000 to 15,000 feet of post-Devonian strata was preserved along its axis from east to west in the Plains, compared to 5,000 to 10,000 feet on the nearby shelf. The axis of the embayment was transverse to the miogeosyncline on the site of the Rocky Mountains and Foothills. Its position lay between latitudes 55 degrees 30 minutes to 56 degrees 15 minutes north, which sector now adjoins the site of the Foothills in the areas of the Peace, Moberly, Pine, and Murray Rivers. The Peace River embayment formed along the north flank of the Peace River arch when this positive cratonic unit foundered and submerged at the close of Devonian time. The continuation of the two palæotectonic units across the site of the Rocky Mountains and Foothills remains to be proven by extensive stratigraphic mapping, but is indicated in part: the Peace River arch, by the absence of Ordovician to Middle Devonian rocks about the Pine Pass (Muller, 1961); the Peace River embayment in the Upper Jurassic-Lower Cretaceous, by sections of the Beaudette Group (Hughes, 1964). The embayment marks an old zone of differential tectonic movement, crustal weakness, and yielding. Movement along this zone persisted in the Laramide revolution; on its north and south, tectonic compression was resolved along different fold trends in the northern and southern Rocky Mountains and Foothills, and the corresponding foreland units reflect different reactions of the basement.

The Inner Foothills underwent continuous close folding over all. It forms part of the orogenic belt. The Outer Foothills, with its discontinuous local folding and lesser deformation, belongs to the orogenic foreland. Close folds of the Inner Foothills pass to the simple fold pattern of the Outer Foothills in the Moberly River-Peace River area. In the Pine River area, the Outer Foothills share the structural trends of the northern Rocky Mountains and Foothills; the Inner Foothills have structural trends which are intermediate or common to those of the northern and southern Rocky Mountains and Foothills.

Structures of the Outer Foothills plunge south-southeast, from the Peace to the Pine River. The plunge affects both anticlines and synclines. The plunge of each anticline exceeds that of its adjacent syncline on the east; values of the plunge range from 65 to 275 feet per mile for anticlines and 35 to 180 feet per mile for synclines, from east to west respectively. The synclinal plunge is correlated with the termination of the Halfway block and the northeast closure of the Alberta syncline. The correlations derive from palimpsest mapping, but only as a generalized solution. Mapping indicates that the relict trace of these major tectonic units of the foreland continue within the Outer Foothills.

A major structural culmination, the Moberly culmination, about the area of Mounts Monteith and Frank Roy shares a common axis with the Fort St. John arch. Oblique folds of minor order also reflect the trace of the common axis across the Outer Foothills in the Moberly River area.

CHAPTER V.—ECONOMIC GEOLOGY

PETROLEUM AND NATURAL GAS

EXPLORATION FOR POTENTIAL RESERVOIRS

The mapping, and descriptions in the text, and the analysis of fold forms in Appendix 1 are of use in exploring for petroleum and natural gas. Few additional comments seem necessary.

Most wells in northeastern British Columbia have been drilled in the Plains. Very little of the Foothills has been tested for oil or gas. The Foothills offer more difficult targets, due to the complexities of folding and the greater thickness of sediments. In addition, access to the drilling-sites is expensive, for there are few roads. However, natural gas in commercial amount has been found in the Foothills in Hunt Sands Sun Falls c-18-G, in the Pine Valley. Small flows of gas were reported in drilling other wells, here.

Few instances of porosity are evident in the Mesozoic rocks of the Pine Valley. The Charlie Lake Formation has porous zones in the subsurface: in sandstones and quartzites, and also in recrystallized quartzites; in dolomites of sub-oolitic and of open drusy texture, which zones are also present in Sun et al Chetwynd 14-20. The quartzites of the Monteith and Monach Formations have occasional and minor discontinuous porosity, but this is not common or widespread. In places, Mesozoic rocks of the Foothills and older rocks were fractured in folding and faulting. The fractures by connecting and draining rocks of low porosity and permeability along the crests of anticlines and along faults can provide other possible reservoirs for oil and gas. Conditions of low porosity are not obvious in exposures and drill cuttings; however, the Charlie Lake, Baldonnel, Beaudette, and Commotion strata contain zones of low porosity with permeability in the subsurface.

The possible reservoirs which may be explored for oil and gas in the subsurface of the map-area can include sandstones and carbonates of Pre-Cambrian, Cambrian, Ordovician, and Silurian age, where present (*see* Muller, 1961); Devonian strata, either reefs and related organic skeletal limestones, dolomitized zones, or sandstones representing the near shore sediments about the extension of the former Peace River arch; coralline and shelly beds of later Palæozoic age; and sandstones and carbonates of the Triassic Grey Beds. These possibilities seem the most important. Others may be found by drilling.

Possible reservoirs of more restricted subsurface distribution may occur in Jurassic and Cretaceous strata in the Outer Foothills and the nearby Plains: the quartzites of the Monteith Formation, the upper quartzite member of the Monach Formation, the basal conglomerate of the Moosebar Formation, and sandstones and conglomerates of the Fort St. John Group—wherever such beds are separated from their outcrop by folding and faulting.

In the Inner Foothills, Palæozoic and Triassic strata contain drilling targets. The presence of extensive flat planar thrusts which may overlie Jurassic and Cretaceous strata in the subsurface here is an unknown factor of the geology and can only be proved or disproved by drilling. The flat planar thrusts have not been found in outcrop. If present in the subsurface, they increase the number of potential reservoirs which can be tested at depth (*see* Appendix 1). The possibilities of flat planar thrusts under the Outer Foothills are examined in Appendix 2.

STRUCTURES

Anticlines receive the primary attention in selecting drilling localities in the Foothills. Data are usually insufficient to deduce positions of stratigraphic traps in the early stages of exploration.

In the Outer Foothills, synclines compose a step pattern of structures. Tectonic movement consisted of the relative depression of the synclines on the east, compression of the long anticlines, and displacements on their faults. The anticlines are concentric folds of low amplitude. Faulting is present; the important faults cut the east limbs. Two components of the anticlines may allow structural traps for oil and gas: the anticlines manifest in the surface structures and corresponding anticlines at depth; and in the footwalls of their faults, subordinate anticlines or upturned edges of the synclines on the east. Closures need to be defined by extensive mapping and to be confirmed by seismic surveys.

On first appearance, complex close folding in angular and concentric forms may discourage exploration in the Inner Foothills. However, specific targets, in Palæozoic and Triassic rocks, may be drilled at less than 6,000 feet depth. Such tests need a detailed appraisal of the structure, and can be directed at anticlines revealed at the surface and those underlying thrust faults. Closure is noted for the Fisher anticline about its culmination, 3 miles northeast of the Pine Valley. It is a large angular fold and forms the west component of the Pine River anticline (anticlinorium). The east component, the Crassier anticline, is thought to obtain closure near the north border of the valley, according to the reports of Beach and Spivak (1944), McKechnie (1955), and the present mapping. However, this structure is greatly affected by faulting, according to the record of the discovery well, Hunt Sands Sun Falls c-18-G.

The forms of anticlines may be obvious in surface mapping, but need to be interpreted for the subsurface according to the data of seismic surveys, geometric projections, and geological concepts. Appendix 1 treats concepts of parallel fold forms and introduces new ideas on the relationships of the folds.

For concentric anticlines of low fold amplitude, interpretations of subsurface geology should anticipate dissimilar structures at surface and depth, and should examine the likelihood of overthrusting, or décollements localized in incompetent formations; the possibilities of overfolding; and the chance of drilling extensive planar thrust sheets. Projections of surface geology to subsurface sections can be unreliable. This is a lesson of the Commotion structure, the only example drilled deeply.

Concentric anticlines of high fold amplitude have fold centres, which are shallow and can be drilled through. In the underlying structures, the drill may enter cuspate folds, cuspate folds modified by faulting, composite folds, or complexes of disharmonic folding and faulting. It need not be assumed that high amplitude concentric folds pass to simpler structures at depth, or that they are simply related to planar décollements. In the Inner Foothills there are three wells drilled on concentric anticlines of high fold amplitude (the Big Boulder, Cairns, and Silver Sands anticlines; *see* following summary). The wells were sited on the west limbs of the anticlines, seemingly to allow for the dip of the axial planes and to intersect the anticlinal crests at depth. According to records for each well, the anticlinal limb is long and steep dipping. Therefore, a great part of the fold mass underlies the fold centre of each concentric anticline. This part of the fold is angular or cuspate, by theory of Appendix 1.

Methods for distinguishing concentric anticlines of low and high fold amplitude are shown in Appendix 1. For those of low amplitude, décollements, or thrusts, and disharmonic structures overlie the fold centres of the anticlines at the surface (or can be so anticipated, from the example of the Commotion anticline; *see* the Tarra and Sarrebrück anticlines, de Sitter, 1956, p. 244). For anticlines of high fold amplitude, the fold centres, obtainable from constructions of Appendix 1, mark the lower limit of the concentric folding expressed at the surface.

Angular folds are not usually tested in petroleum exploration. Their forms can persist to great depth, but underlying structures, thrust faults, décollements, or perhaps concentric folds, which may be present, are not predictable from surface mapping alone. Fault breaks along the axial plane are common. It is possible that some were opened by tension, a condition which would affect the emplacement of oil and gas.

Most anticlines in the Foothills are long. Closure is unmapped or ill defined for many of them. In the Outer Foothills, the prevailing south-southeast plunge does not rule out culminations. Also, detailed mapping may reveal subordinate flexures, distributed in echelon along the anticlines.

In areas of folded structures like the Foothills, several wells may be drilled before the economic potential of anticlines can be appreciated. The early wells permit reinterpretations of the subsurface structure for later drilling. For this reason, previously drilled structures can still be considered favourable for exploration. The need for reliable seismic surveys is emphasized, for combinations of several fold forms, faulting, and disharmonic fold patterns can be expected in the subsurface.

WELLS DRILLED FOR OIL AND GAS

Seven exploratory wells have been drilled to date in the Pine Valley. Of these, five entered complex structures or close folded anticlines, one well was shallow, and the seventh drilled flat-lying beds near the border of the Foothills and the Plains. The following note summarizes the economic geology of the well sections.*

Sun et al Chetwynd 14-20-77-23

Location: Legal Subdivision 14, Section 20, Township 77, Range 23, west of the 6th Meridian.

Elevation: 2,669 feet K.B. *Total depth:* 9,403 feet.

Status: Abandoned, 1963 (drilled to 4,207 feet as Herkimer Pure Chetwynd No. 1, and abandoned; deepened to 8,360 feet as O'Sullivan-Pure Chetwynd No. 1; remaining footage drilled under Sun Oil Company).

Strata (depths to tops of stratigraphic divisions in feet): Spudded in Dunvegan Formation; Fort St. John Group—Cruiser Formation 1,210—Shales with Sandstones 1,842—Hasler Formation 2,602—Commotion Formation 3,387—Moosebar Formation 4,804; Crassier Group—Gething Formation 5,887—Dresser Formation 6,400—Brenot Formation 7,070; Chetwynd Beds 7,375; Beaudette Group 7,620; Fernie Group—Transition Beds 8,308—Middle Shales 8,405—Nordegg Beds 8,718; Schooler Creek Group—Grey Beds 8,815 (Baldonnel Formation 8,815—Charlie Lake Formation 9,240).

Remarks: The Chetwynd well was drilled on the Wabi structure, near the east boundary of the Foothills. The well section is unfaulted and the strata are flat lying (Fig. 2).

The Fort St. John section is comparable with those of surface sections to the west in the Outer Foothills, but there are some differences in the formational divisions. Below the Cruiser shales, the unit, Shales with Sandstones, containing four arenaceous members (1,800 to 1,842, 2,160 to 2,260, 2,315 to 2,460, and 2,516 to 2,628 feet), occupies a stratigraphic position corresponding to the Goodrich Formation. The same distribution of four sandstone members lies in the interval, 548 to 1,485 feet, of the Anglo Bralsaman Little Prairie No. 1 well to the west; the upper part of the Fort St. John Group is similar in both wells. The Hasler, Commotion, and Moosebar Formations show little change from their outcrop sections (*see* Subsurface Stratigraphy, Chapter III). The Crassier Group consists of coal measures, and its three formations are distinct in geophysical and lithological records. A unit of marine and non-marine strata, named the Chetwynd Beds, separates the Crassier and Beaudette Groups. The Beaudette Group contains no obvious divisions of formational rank. Divisions of the Fernie Group resemble those of the Foothills. The Triassic section, the Grey Beds containing the Baldonnel and part of the Charlie Lake Formation, is similar to that in the subsurface of the Plains.

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* Complete records of the wells, including geophysical logs and drill cuttings, are available for study at the Department of Mines and Petroleum Resources, Victoria; drill cuttings and cores are similarly available at the Charlie Lake Field Office; and a third set of cuttings at the Geological Survey of Canada, Calgary.

In drilling the well, gas shows were reported for the intervals 3,390 to 3,457 (Commotion Formation), 3,594 to 3,662 (Commotion Formation), and 4,124 to 4,207 (Commotion Formation). These horizons issued gas on drill-stem tests, but at rates too small to measure. The well flowed at 750 thousand cubic feet of gas per day, on open-hole test, measured through a ¼-inch orifice of a 2-inch critical flow prover. The operators noted about 123 feet of porosity and oil staining in the Grey Beds below 8,800 feet depth; this interval failed to produce after perforation and acid treatment.

Anglo Bralsaman Little Prairie No. 1

Location: Legal Subdivision 11, Section 21, Township 77, Range 24, west of the 6th Meridian.

Elevation: 2,106 feet K.B. *Total depth:* 1,830 feet.

Status: Abandoned, 1955.

Strata: Fort St. John Group—drill entered the Cruiser Formation in the interval 0 to 121 feet, Shales with Sandstones 548—Hasler Formation 1,465.

Remarks: The well was drilled on the Chetwynd anticline. Figures 2 and 3 show its site in the footwall of a thrust fault cutting the west limb of the anticline. Near the well the Cruiser-Dunvegan boundary lies at elevations between 2,250 and 2,500 feet; the Dunvegan strata dip southwest 5 degrees to +10 degrees. Farther west, across a small wooded valley, the Cruiser-Dunvegan boundary is at about 2,600 feet elevation, and falls on the southwest to the axis of the Bissett syncline. The well provides little information on structure; self-potential and resistivity logs were run; no cores were taken; there is no record of dips in the well section.

The upper part of the Fort St. John Group is mostly shales. The interval 548 to 1,465 feet, assigned to the subsurface unit, Shales with Sandstones, contains four arenaceous members in the intervals (i) 1,340 to 1,465, (ii) 1,110 to 1,220, (iii) 890 to 1,060, and (iv) 548 to 620. Shales with lesser siltstones and few sandstones separate the arenaceous members and their individual sandstone sets.

Anglo Bralsaman Little Prairie No. 1 bottomed in the upper part of the Hasler Formation. No oil or gas shows were found, and no tests were run.

British Columbia Government Pine River No. 1

Location: Quarter d, Unit 23, Block E, of National Topographic Series Map 93-P-12 (*see also* description of location, Appendix 3).

Elevation: 2,010 feet K.B. *Total depth:* 6,941 feet.

Status: Abandoned.

Strata: Spudded in Quaternary drift; Fort St. John Group—Moosebar Formation and ?lowermost Commotion beds 1,081; Crassier Group—Gething Formation and upper part of Crassier Group 2,424.

Remarks: The well tested the Commotion anticline. Drilling was abandoned in complex structures without shows of oil or gas. (*See* Chapter IV and Appendix 3 for full account.)

Hunt Sands Sun Falls c-18-G

Location: Quarter c, Unit 18, Block G, National Topographic Series Map 93-O-9.

Elevation: 2,974 feet K.B. *Total depth:* 10,590 feet.

Status: Capped Triassic gas well, 1962.

Strata: Spudded in Crassier Group; Fernie Group—Middle Shales 6,847—Nordegg Beds 8,910; Schooler Creek Group—Pardonet Formation 9,193—Grey Beds 9,436 (Baldonnel Formation 9,436); Fernie Group—Nordegg Beds 9,528; Schooler Creek Group—Pardonet Formation 9,702, Grey Beds 9,944 (Baldonnel Formation 9,944).

Remarks: The well was drilled on the Crassier anticline, the eastern component of the Pine River anticline or anticlinorium (Fig. 2). The surface and subsurface structures are unlike. The well section is complicated by folds and faults; Beaudette strata are excluded; and vertical intercepts of the Crassier and Fernie beds far exceed those due to projecting the trace of surface structures. The Crassier coal measures persist down to 6,850 feet, with little to distinguish the separate formations in the drill cuttings and geophysical records (gamma ray and sonic logs). The entire Crassier section contains steep dips, folds, and probably thrust faults. For the complex lithology of the coal measures, the disturbed attitudes of the beds mask the register and contrast of the formations on the logs. Similarly, in drill cuttings, criteria to differentiate the formations, the sand/shale ratios and the distribution of coals and sandstones over wide stratigraphic intervals are also rendered uncertain. Coal measures of the Brenot and Gething type predominate in the cuttings. The Dresser Formation, with its full assemblage of coarser-grained sandstones and grits, appears to be missing or only partly represented. The bulk of cuttings in the interval 4,980 to 6,120 feet consists of coal, a condition which indicates abnormal thickening of Gething (or Crassier) beds by steep dips, folding about a structural axis, or thrust faulting. The Crassier Group rests on the Middle Shales, which unit of the Fernie Group occupies the interval 6,847 to 8,910 feet. There are few lithological features of this unit. Three zones of soft fissile shales, with abnormal, low velocity bands recorded on the sonic log, occur in the intervals 7,012 to 7,064 feet, 7,530 to 7,640 feet, and 7,918 to 7,975 feet. They are regarded as shear or fracture zones, and may mark fault breaks. The interval 8,910 to 9,193 feet contains black shales and calcareous shales, argillaceous limestones, and minor chert, and is assigned to the Nordegg; this unit contains bands of relatively high radioactivity. The Nordegg overlies Pardonet and Baldonnel beds in normal order, and this sequence is repeated below 9,528 feet. Drilling ended in the Baldonnel Formation.

The Crassier anticline at the surface is broken by a fault, or fault zone, trending south 50 to 55 degrees east and lying 1,750 feet northeast of the well-site. Details of the fault break are obscure, and it is marked by a set of minor folds or dragfolds, and the discordance between the simple upturning of strata bordering the Hulcross syncline on the northeast, and the complex and faulted structures in the rest of the Pine River anticline to the southwest. The Dresser Formation occupies the crest of the anticline, and the Dresser and Gething Formations outcrop on the northeast of the fault zone. The structure of the well section can be expressed as follows:—

(*a*) Folding and faulting of the Crassier beds from surface to 6,847 feet.

(*b*) The contact of the Fernie and Crassier beds at 6,847 feet.

(*c*) The section of Fernie strata in the interval 6,847 to 8,910 feet—a long intercept due to steep dips or folding, with or without thrust faulting.

(*d*) Faulting at 9,528 feet, and repetition of the Nordegg and Schooler Creek beds below this point.

Structures under (*a*) and (*c*) have been noted; they probably include thrust faults as well as folds. Several explanations for the contact of Fernie and Crassier beds (*b*) are discussed: The removal of Beaudette strata by pre-Crassier erosion at an unconformity seems unlikely. It represents a special condition, unknown elsewhere in the map-area. Calcite vein material in the cuttings, about the Fernie-Crassier contact, 6,840 to 6,880 feet, points to faulting here. The contact may mark a thrust or a normal fault. For a thrust, the condition by which Beaudette strata were excluded in the movement is necessary; thus the possibility of a thrust or a folded thrust with a step or sigmoid traces which cut down into Beaudette strata on one side of the well. A normal fault allows a simpler explanation for the Fernie-Crassier contact. The thrust fault (*d*), repeating the Nordegg and parts of the Triassic, is shown clearly by the gamma ray and sonic logs. A continuous dipmeter log is available for parts of the well section. It requires detailed interpretations for the most part, as many of the dip records are rated " poor " or show a scatter of values. Interpretations are best left to the individual geologist.

The Grey Beds yielded a steady gas flow, reported to be 3.6 million cubic feet per day on drill-stem test, in the interval 9,415 to 9,524 feet. The well was completed as a potential gas producer with a declared productive interval of 9,430 to 9,490 feet in the Triassic. The operators reported a flow of 13.8 million cubic feet of gas per day, absolute open-flow potential, in final tests. Their analyses of the gas may be summarized: methane, 38 to 42 per cent; carbon dioxide, 17 per cent; hydrogen sulphide, 38 to 42 per cent. The large proportions of carbon dioxide and hydrogen sulphide are noteworthy. The well should be expected to produce much sulphur.

Hunt Sands Sun Boulder b-74-D

Location: Quarter b, Unit 74, Block D, National Topographic Series Map 93-O-9.

Elevation: 2,200 feet K.B. *Total depth:* 4,820 feet.

Status: Abandoned, 1963.

Strata: Spudded in sands and gravels (Quaternary); Beaudette Group—Monteith Formation 1,460.

Remarks: The well was drilled on the Big Boulder anticline (Fig. 2). The drill entered the structure near a sharp flexure of beds which marks a separation between the west limb of the anticline and its broad flattened crest. The Monteith Formation is represented in the drill cuttings by very fine- to medium-grained sandstones, and quartzitic sandstones, minor siltstones, and few intervals of black shales and silty shales, from bedrock to total depth.

Monteith beds outcrop in the core of the anticline near the well-site. The surface and subsurface geology reveals a vertical intercept of +4,800 feet for the Monteith Formation, that is, about three times the stratigraphic thickness of the formation. The long intercept may be due to thrust faulting, or, more probably, to steep dips of the order of 70 degrees or more. At the surface, Beattie Peaks and Monach beds dip 40 to 70 degrees southwest in the west limb of the anticline. The anticline is a high amplitude, concentric box fold. Quite possibly the Monteith beds in its west limb are partly infolded at depth—a simulation of diapiric folding. Such folding may be masked by minor disharmonic structures in Beattie Peaks, Monach, and Crassier beds on the west limb of the anticline. Cores in the interval 4,725 to 4,820 feet showed dips between 70 and 80 degrees (company report).

The core of the Big Boulder anticline was not drilled. No oil was found; gas was present, but the amount was insignificant. A drill-stem test, 3,520 to 3,717 feet, recovered water and a blow of gas too small to measure (company report).

Triad Bush Mountain b-23-A (1)

Location: Quarter b, Unit 23, Block A, National Topographic Series Map 93-O-10.

Elevation: 2,358 feet K.B. *Total depth:* 10,614 feet.

Status: Abandoned, 1959.

Strata: Spudded in sands and gravels (Quaternary); Schooler Creek Group—Grey Beds ca. 485 (Baldonnel Formation ca. 485—Charlie Lake Formation 530—Halfway Formation 1,150)—older Triassic and ?Palæozoic 1,680.

Remarks: The well was sited on the west limb of the Cairns anticline (Fig. 2). The Grey Beds and Pardonet outcrop in the core of the anticline and dip about 50 degrees southwest in its west limb. The anticline is a close concentric fold of high amplitude, and its axis lies 2,500 feet northeast of the well-site. The west limb of the anticline contains a subordinate fold developed in the Pardonet beds at the surface (elevation 2,800 feet). A structural break lies at the base of the fold, and it may continue southwestward, intercepting the well section. The Grey Beds can be traced to about 1,680 feet, the lower boundary being drawn in a transition to the Dark Siltstones. The underlying Triassic interval is left unclassified. The vertical intercept of Middle and Lower Triassic strata shows that a long interval was drilled in the west limb of the Cairns anticline. Dips reported from the cores are 25 to 45 degrees for places in the interval 5,248 to 7,911 feet, and 80 to +80 degrees for places in the interval 8,797 to 10,614 feet. No oil or gas was found in drill-stem tests.

Triad BP Bush Mountain a-15-A

Location: Quarter a, Unit 15, Block A, National Topographic Series Map 93-O-10.

Elevation: 2,312 feet K.B. *Total depth:* 10,967 feet.

Status: Abandoned, 1959.

Strata: Spudded in gravels (Quaternary); Schooler Creek Group—Grey Beds 90 (Baldonnel Formation 90, Charlie Lake Formation 630)—older Triassic and Palæozoic 1,550.

Remarks: The geology of Triad Bush Mountain a-15-A, on the Silver Sands anticline, resembles that of the earlier b-23-A (1) well (Fig. 2). The Silver Sands anticline is a close concentric fold of high amplitude. The Charlie Lake, Baldonnel, and Pardonet Formations outcrop in the core of the anticline and dip 40 to 65 degrees southwest in the west limb; the west limb is unfaulted at the surface. The well-site (a-15-A) is near an exposure of Baldonnel limestones with *Spiriferina*, in the west limb of the anticline. The drill entered the lower part of the Baldonnel at 90 feet and the Charlie Lake at 630 feet depth. Drill cuttings from 1,550 to 2,960 feet are missing. The remainder of the well section is not examined in detail. Dips ranging from 20 degrees to 80 degrees were reported from cores taken at places from 3,636 to 10,908 feet. Most or all of the well section kept within the west limb of the Silver Sands anticline. A small gas blow, of 15 thousand cubic feet per day, was reported for a drill-stem test in the interval 9,482 to 9,732 feet, in the Palæozoic. There were no other oil or gas shows obtained in tests.

COAL

The Crassier Group has important coal reserves in the Pine Valley. Some were investigated and reported by McKechnie (1955), whose estimates were 9 million short tons for the Noman Creek area, and 23.8 million short tons for the Willow Creek area (estimates based on mineable seams 4 feet and more thick). Other outcrops of the Crassier Group were not included by McKechnie, namely, parts of the Crassier and Fisher Creek areas, and the areas west of Mount Bickford and Beaudette Creek. Coal-mining on a commercial scale has not been attempted in the Pine Valley.

Most of the mineable seams occur in the upper half of the Crassier Group, that is, in the upper part of the Dresser Formation, and within the Gething Formation, in outcrops east of Mount Bickford and Beaudette Creek. Here, McKechnie (1955) named 10 seams which are more than 5 feet thick, and which are common to the uppermost 1,000 feet of the Crassier Group. The thickness of the designated seams may range to more than 10 feet, inclusive of bone and non-coaly partings. Seams vary in thickness from place to place; splitting of the coal seams appears to be common. Seams in the lower part of the Dresser Formation and within the Brenot Formation are thin; none are found to exceed a thickness of 4 feet, which can be regarded as the least thickness for mining. These lower parts of the Crassier Group have received little attention in coal surveys (Spivak, 1944; Mathews, 1947; McKechnie, 1955). The Crassier coal measures, in outcrops west of Mount Bickford and Beaudette Creek, have not been evaluated. Much of the upper part of the Crassier Group has been removed by post-Cretaceous erosion, here; many of the more valuable seams are likely to be missing, or have a small area of subcrop and distribution; also there are many small-scale folds and faults. Folding and faulting, on small and large scales, disturbed the Crassier coal measures in the Pine Valley. The deformation, though severe locally, does not prohibit mining, and may favour surface stripping. However, deep weathering has spoiled the quality of the coals in many places; analyses of samples from natural exposures are found to be unreliable. To evaluate reserves it is necessary to investigate coal seams by trenching, stripping, and diamond drilling.

In the upper part of the Crassier Group, the coals are of good quality and include a large proportion of medium- to low-volatile bituminous rank coal. Many show low ash and sulphur analyses, and some are reported to have good coking qualities (McKechnie, 1955). The high calorific value of the coals is well known.

There is little demand for the coal. The market available now in the Peace River District is supplied by the Gething and Reschke mines near Hudson Hope. Petroleum and natural gas, locally produced and refined, are expected to provide fuels to this market for some time to come.

Coal seams found in the non-marine and uppermost member of the Commotion Formation (Member (iv)) are less than 2 feet thick, and on this account are unprofitable for mining. No analyses are known. There is little prospect of mineable coal seams in the Dunvegan Formation.

APPENDIX 1

PARALLEL FOLD FORMS

REVIEW AND DEFINITIONS

Folding in the Foothills is of parallel type, and the fold forms are concentric and non-concentric.

Such forms are analysed from the mapping of the Pine Valley.* This treatment forms the basis of the following discussions, which include theories of the development of parallel fold forms deduced from geological structures. It is advisable to review and define concepts of parallel fold forms, according to a classification introduced here.

Parallel folds are defined as folds in which strata maintain uniform thickness and bedding planes remain parallel in folding. Three classes of parallel folds are recognized—concentric, cuspate, lambdate.† They are distinguished from the right sections of folds (Fig. 8). Also, a general class—angular—contains the cuspate and lambdate folds as explained in the following definitions.

CONCENTRIC FOLDS

In concentric folds, strata are bent in arcuate form, and their traces compose a set of parallel non-intersecting curves in right sections. The folds have a recognizable fold centre.

Two types of concentric folds, those of high and low fold amplitude respectively, have been discovered from the field work in the Pine Valley.

The terms concentric and parallel are often used as synonyms. Concentric folds are parallel, but the reverse is not necessarily true. Mertie (1940, p. 1111) discussed this misplaced synonymy at length.

Concentric folds are represented in right sections by arcs of circles or by involutes. The choice is significant, for it influences our concept of the fold centre.

Arcs of circles representing folded strata, in the methods of Hewett (1920) and Busk (1929), are constructed from field observations; the plotted dips form tangents to the circular arcs. A single fold may be composed of one or a series of arcs; adjacent arcs are joined by a common tangent of finite or negligible length; all beds are concentric in each arc composing the fold. This method is shown in Figure 11.

Concentric folds are also represented by the "general method of evolutes and involutes" proposed by Mertie (1940). In this method, folded strata are shown by involutes, and the involutes plotted from an evolute which is constructed from the normals of the dips. Involutes drawn from any evolute compose a family of curves which are parallel. Normals to the involute are tangents to the evolute. Mertie stated that this method records the true curvature of the fold. Later (1947, 1948), he published other special constructions for representing parallel folds by involutes. The fold of Figure 10 is drawn by the general method of Mertie (1940).

The concept of fold centres in concentric folds is important in a geometric and mechanical sense. In the construction of Hewett (1920) and Busk (1929), the fold

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[*] Also, a few examples from the Moberly and Peace River areas are presented for completeness and illustration.

[†] The classes of parallel folds proposed here have been worked out independently of a contemporary classification by Hills (1963). (Reference: Hughes, J. E.—" Peace and Pine River Foothills: Structures and Tectonics," thesis, McGill University, 1963.) The two classifications are alike. However, the writer has changed some names already proposed by Hills, with the intention of improving the terminology of parallel folds. The definitions in the text express the writer's view.

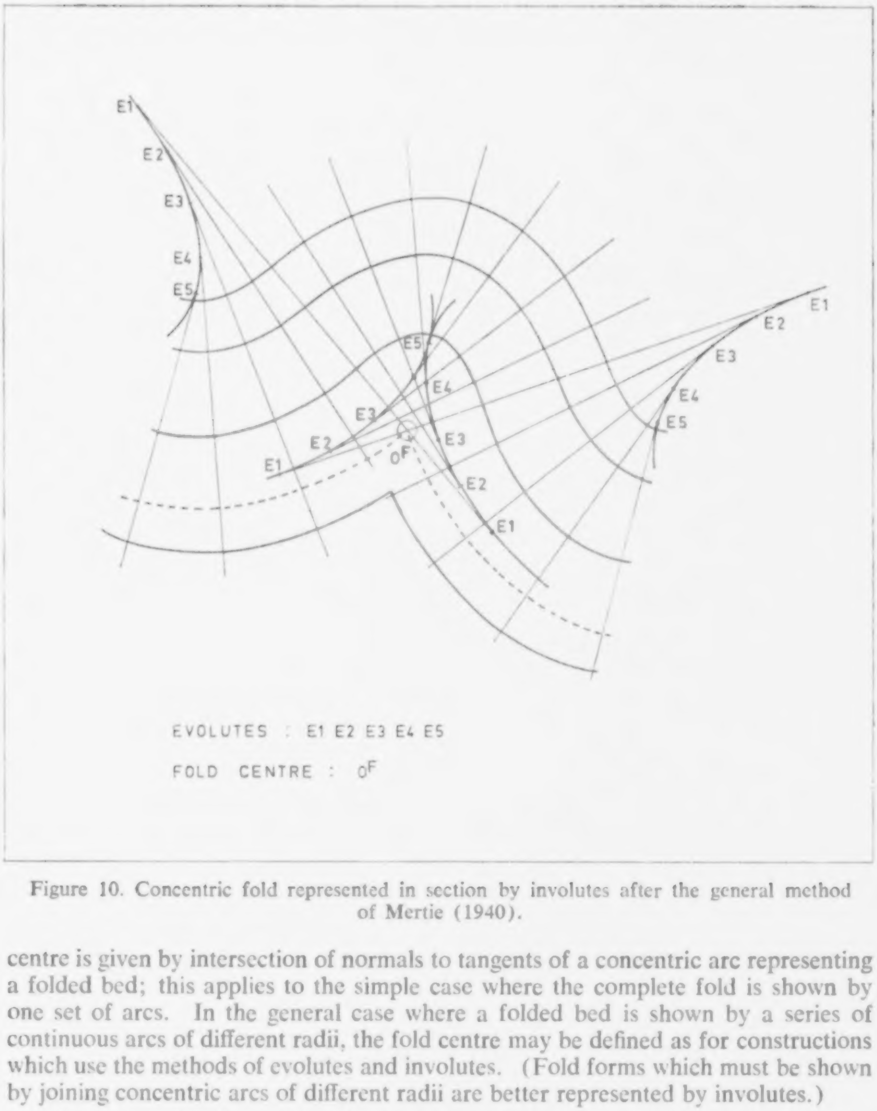

Figure 10. Concentric fold represented in section by involutes after the general method of Mertie (1940).

centre is given by intersection of normals to tangents of a concentric arc representing a folded bed; this applies to the simple case where the complete fold is shown by one set of arcs. In the general case where a folded bed is shown by a series of continuous arcs of different radii, the fold centre may be defined as for constructions which use the methods of evolutes and involutes. (Fold forms which must be shown by joining concentric arcs of different radii are better represented by involutes.)

In the constructions of Mertie the fold centre is less easily defined, as involutes representing strata have no stationary point of centre.

A general definition of the fold centre is then given. In anticlines, arc(s) or involute(s), convex upward, represent strata in the major sector of the fold between limbs. The fold centre for anticlines is the point at which the convex arc(s) or involute(s) are infinitely small or become excluded (Figs. 10 and 11). Concave replaces convex in defining the fold centres of synclines.

The fold centre separates two parts of the fold: in anticlines for example—the upper part in which the folding was concentric, and the lower part in the core where the folding was non-concentric. The anticlinal core then contains a cuspate fold, or a complex of multiple folds and faults, and disharmonic structure.

CUSPATE FOLDS

They are parallel folds without fold centres. Traces of the limbs are arcuate to almost straight in right section, and can be represented by intersecting circular arcs or involutes. Folding is by angular bending of strata across the axial plane. The apical angle, measured between the limbs across the axial plane, varies throughout the fold, decreasing downwards in anticlines. The traces of beds shown in section are incongruent when superimposed. The folds are regarded as non-similar.* Beds have equal thickness in the limbs, except for their intercept on the axial plane.

The writer is obliged to Hills (1963) for the word cuspate. It is a general term in mathematics and architecture.

LAMBDATE FOLDS

This class includes all parallel folds with straight limbs. Beds have uniform thickness in the fold, except for their intercept on the axial plane. The angle between the limbs is the same at all levels. Traces of folded beds are congruent when superimposed. Lambdate folds are similar. They do not appear to be specially related to concentric folds.

The lambdate fold class includes the accordion folds of de Sitter (1956, p. 216), the chevron folds of Billings (1942, p. 41), and probably the chevron folds of de Sitter (1956, p. 222). Both the accordion and angular folds of Hills (1963) are brought under this class. The accordion fold is recognized as a serial pattern of the lambdate fold form. It may represent the general or, perhaps, the only type of its class, but this view must stand the test of future studies. De Sitter thought the accordion fold to represent a kind of cleavage folding.

The word lambdate is from the Greek letter △. It is unambiguous and descriptive of folds with straight limbs.

ANGULAR FOLDS

This general term includes all non-concentric parallel folds marked by a sharp angular bending of strata across the axial plane; that is, the cuspate and lambdate folds.

The term is necessary where the distinction of cuspate and lambdate forms is impracticable due to insufficient field evidence and modifications of the fold form. Cuspate folds with slight curvature may be confused with lambdate forms where exposures are poor. Simple and ideal fold forms are modified by faulting and the development of subsidiary folds. Also, the term angular covers the logical possibility of parallel folds with one straight and one curved limb, a condition which is almost realized in the field example of Plate IV.

REMARKS

Classifications of folds depend on abstractions and are referred to ideal geometric models (Mertie, 1959, p. 91). All parallel folds are reducible to the basic forms concentric, cuspate, and lambdate, whether they have cylindrical, prismatic, periclinal, conical, or pyramidal shapes in three dimensions.

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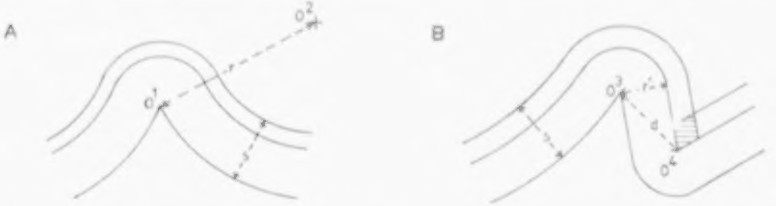
[*] The term similar, as generally used, fails to decide whether the form of a given fold or whether the fold form continues unchanged at depth—two different conditions. The cuspate fold particularly indicates this ambiguity. We owe the name similar to van Hise (1896, p. 599), who gave it for non-parallel folds which "can persist at depth" because of the geometry of their forms. He illustrated the ideal or idealized similar fold by a set of congruent curves. There was later a tendency to accept similar and non-parallel as synonymous terms. Mertie (1959) recognized similar folds to form one class of non-parallel folds, basing this restricted concept on a mathematical definition of similar surfaces for congruent and incongruent fold forms.

Differential thickening from limbs to axes may occur in parallel folds, but this is small and insufficient to invalidate the general principle of parallel folding. Also thickening, caused by faulting, and subsidiary, disharmonic folding, may be found in any part of these folds. The three are concomitants of parallel folding—a fact which has been long recognized, at least, tacitly. Disharmonic folding itself represents various combinations of faulting, concentric, angular, and non-parallel folding, the last usually taking the form of restricted and local cleavage folding in incompetent beds.

The following generalities apply to folds in the Foothills of the Peace, Moberly, and Pine River areas. Major anticlines and synclines may be simple or compound (=anticlinal and synclinal belts=anticlinoria and synclinoria). These descriptive terms are used, but they do not hold any systematic value, for instance compound anticlines may pass to simple anticlines, as in the Butler Ridge-Portage Mountain structure of the Peace River area. The anticlinal belts are long, but their forms and the developments of component anticlines vary along their lengths. Most anticlines are thrust faulted on the northeast limbs; the folding and thrust faulting occurred together.

Differences of the parallel fold forms imply different mechanisms and conditions of deformation. These relationships are only partly understood. The following discussions on parallel folding concern the gross morphology of the folds in terms of the classification already expressed. The theories of folding mechanisms, included in the discussions, concern the development of fold forms and their structural relationship. They are based on evidence obtained in field mapping and illustrated by local examples. This treatment partly complements theoretical and experimental studies of parallel folding. For these, reference can be made to Ickes' (1923) distinction of concentric folding by flexure with interlayer slip and by simple bending; Keunen and de Sitter's (1938) experimental verification of layer-parallel shear in concentric folding; Goguel's (1948), Biot's (1961), and Ramberg's (1963) theories relating compressive stress, rock viscosity, elastic moduli, and time in the folding of stratified viscoelastic media under different assumptions of layer-parallel shear and interlayer slip.

Theories of folding introduced in the following text assume representation of concentric and cuspate folds by circular arcs. Models of this type allow a first approximation of natural conditions.



Simple concentric folding around centres O^1 and O^2. The bed thickness s is less than the distance r between the fold centres.

Concentric folding around centres O^3 and O^4 for strata of thickness s. The thickness s is greater than r' the distance between the fold centres measured perpendicular to the bedding. The approximate overlap of volumes for bed thickness s is shown by horizontal shading.

Figure 11. Geometric relationships in concentric folding.

CONCENTRIC FOLDING: GENERALITIES AND APPLICATIONS IN THE FOOTHILLS

In concentric folding, beds under compression are bent about fold centres. The radii of folding of the beds decreases as compression and folding develops. In the folding the volumes of the strata remain sensibly unchanged. Change of volume due to the internal strain of the rocks and the reorientation and recrystallization of the sedimentary fabric can be considered negligible. Concentric folding requires a partial décollement at depth and some non-concentric deformation in the core, below the fold centre. Deformation and movement in the core may be provided by flowage; a complex of multiple folding, faulting, and shear; or by simple cuspate folding, at least in part.

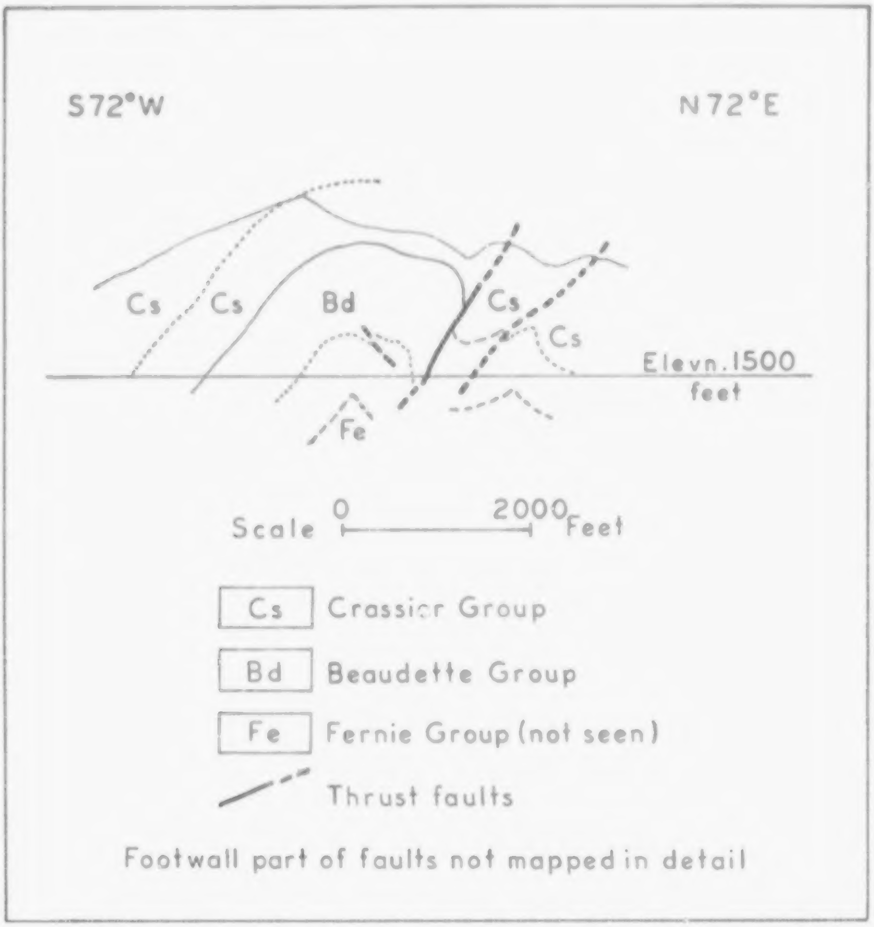

S72°W
N72°E
Cs
Cs
Bd
Cs
Cs
Elevn. 1500 feet
Fe
Scale 0 2000 Feet
Cs Crassier Group
Bd Beaudette Group
Fe Fernie Group (not seen)
Thrust faults
Footwall part of faults not mapped in detail

Figure 12. Section of the Portage Mountain anticline at Grant Knob, Peace River Canyon.

In stages of concentric folding shown by Figure 11,

s—the thickness of strata,

d—the distance between fold centres,

r or r'—the distance between adjacent fold centres measured perpendicular to the bedding,

l—the length of a stratum in section, in any given fold sector.

For Figure 11-A the concentric folding is semi-circular and gives a series of simple folds for a group of strata such that s is less than r or r'. A limit of concentric folding in geometric sense is given by the condition $s = r$ or r'.

Where s is greater than r or r', and the ratio r/l is maintained for each of the beds around a given fold centre, an interference of volume occurs in the fold limb (Fig. 11-B).

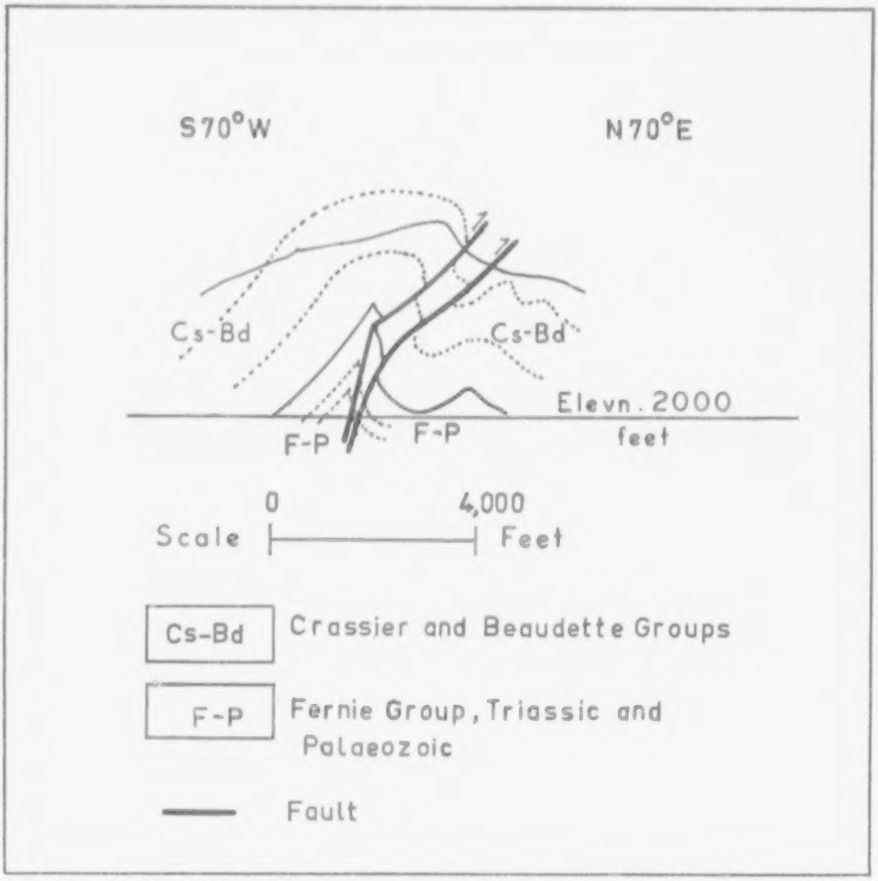

S70°W
N70°E
Cs-Bd
Cs-Bd
Elevn. 2000 feet
F-P
F-P
0
4,000
Scale
Feet
Cs-Bd
Crassier and Beaudette Groups
F-P
Fernie Group, Triassic and Palaeozoic
Fault

Figure 13. Section of the Mount Gething-Stott Creek structure after Beach and Spivak (1944).

The concentric folding then becomes discontinuous. This condition obtains in close folding. The excess volume developed in the limb may be accommodated in several ways: (1) by major reverse faulting across the steep limb in thrust-faulted anticlines; (2) by subsidiary folding, or faulting directed toward the anticlinal axis; (3) by multiple subsidiary folding within the adjacent syncline. Reverse (thrust) faulting is common in concentric anticlines at early or late stages of their development; the Portage Mountain anticline illustrates this condition (Fig. 12). Subsidiary folds and faults are frequent in concentric folding, as in the Commotion anticline and in the Coyote Creek syncline against the Big Boulder anticline (Fig. 2). The above discussion expands a similar treatment of volume and structural relationships in concentric folding by de Sitter (1956, p. 241).

In the Foothills, sequences with thick units of sandstones and conglomerate tend to fold concentrically about cores of non-concentrically deformed shales and other weak beds. A relation of the structures to the competency of the strata is observable in mapping. The most incompetent beds are the shales of the Fernie Group. Other weak beds include parts of the Triassic, parts of the Beattie Peaks Formation, the coal measures of the Crassier Group, and the shales of the Fort St. John Group. In the Foothills much of the concentric folding was made possible by shales of the Triassic and of the Fernie Group. The fold centres of several anticlines are traced to them from cross-sections at the surface. The major fold pattern is then expressed in the competent beds of the Schooler Creek and Beaudette Groups. Non-concentric folding and fault movement can be shown in beds which underlie the concentrically folded structures, or is indicated. Examples include the Mount Gething-Stott Creek structure (Fig. 13) from Beach and Spivak (1944), the Commotion anticline (Chapter IV), and the Le Moray syncline (Fig. 3).

The mechanism of concentric folding, described by de Sitter (1956, pp. 196–199), is reviewed for its importance in the later discussions on folding. In the folding of an anticline, the radii of folding strata decrease and the fold centre moves up in stratigraphic level. Concurrently beds in the core of the fold below the fold centre undergo non-concentric deformation. The fold centre rises as fold compression continues, and it is followed by the spread of non-concentric deformation. It is a common concept in structural geology that concentric folds require non-concentric deformation and décollements at their base. The writer subscribes to this view, and adds also the concept of cuspate folding, underlying, and replacing concentric folds.

CONCENTRICALLY FOLDED ANTICLINES: TYPES OF LOW AND HIGH FOLD AMPLITUDE

The concentric anticlines of the Foothills include two types—those of low and high fold amplitudes. Respective examples are the Commotion structure (Fig. 2) and the Monteith anticline (Fig. 14). Low amplitude folds have limbs of slight or moderate dips and open fold form. High amplitude folds are close folded. The arcs of the middle or axial parts may have the same curvature* in both types (Fig. 8).

The two fold types are separated by means of indices derived from their fold forms: *the index of fold amplitude* and *index of anticlinal amplitude*. The indices are explained in the next section. In general, different amplitudes of concentric anticlines indicate differences in compression, fold habit and thrust faulting, and dependence on different distributions of competent and incompetent strata.

Thrust faulting commonly develops in the folding of the anticlines. Folding and thrust faulting combined relieve compression and provide lateral contraction in a given structure. Each may replace or substitute for the other.

Interreplacement of folding and faulting is not simple or continuous in the growth of these anticlines; two distinct habits of fold development are recognized. They are divergent and alternative developments arising at an early stage of growth. They result in low and high amplitude forms, without intermediary types.

The anticlines of low amplitude are broad. The Commotion anticline is 2 miles wide and has a simple form. The Chetwynd (Little Prairie) anticline, 3 to

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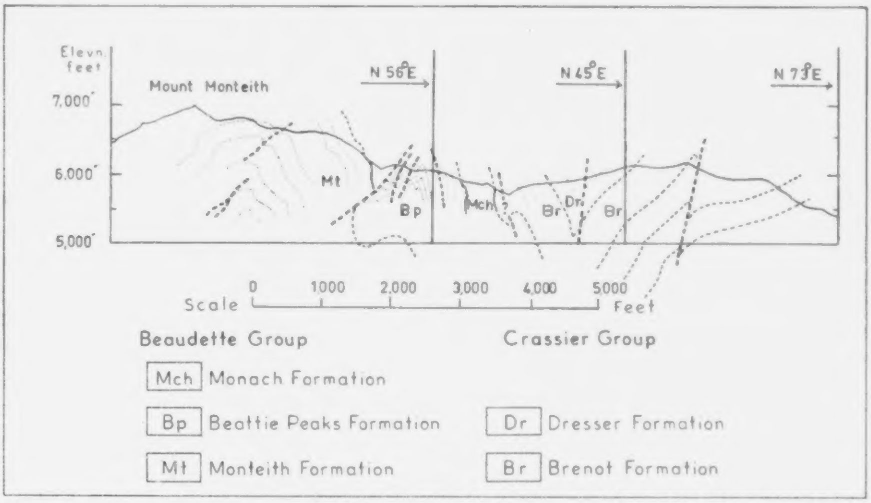
* Curvature refers to the factor 1/R where R is the radius of the curve.

4 miles wide, is compound. It is modified by reverse faulting in part of the west limb. The Hudson Hope anticline in the west margin of the Plains resembles it. The oblique anticlines of the Outer Foothills have low fold amplitudes also, for example, the Hulcross anticline (Wickenden and Shaw, 1943, p. 13). Anticlines of low amplitude are found in the Outer Foothills and the adjoining Plains. They are developed in beds of the Fort St. John Group and in the Dunvegan Formation, the upper part of the stratigraphic succession in the Foothills.



Elevn. feet
7,000'
6,000'
5,000'
Mount Monteith
N 56°E
N 45°E
N 73°E
Mt
Bp
Mch
?
Br
Dr
Br
Scale
0
1,000
2,000
3,000
4,000
5000
Feet
Beaudette Group
Crassier Group
Mch Monach Formation
Bp Beattie Peaks Formation
Mt Monteith Formation
Dr Dresser Formation
Br Brenot Formation

Figure 14. Geological section, Mount Monteith.

Anticlines of high amplitude occur in the Inner and Outer Foothills, in Bullhead and Triassic strata. The Cairns and Beattie Peaks anticlines have subcircular forms. Box-form anticlines have a distinct inflection of the dips at the junction of the limbs and the flattened axial parts. They represent a more advanced condition of compression. The box forms appear to be common, and were noted by Beach and Spivak (1944) and Mathews (1947). Examples include the component anticlines of the Butler Ridge structure (Beach and Spivak, 1944, p. 12), the Big Boulder anticline, and the Silver Sands anticline in the Pine Valley.

DISTINCTION BETWEEN HIGH AND LOW AMPLITUDE CONCENTRIC FOLDS

Concentric anticlines of high and low fold amplitudes can be separated by two derived indices: *the index of fold amplitude* and *the index of anticlinal amplitude*.

These indices can be shown first for simplified models of concentric folds, in which anticlines and adjacent synclines are represented by concentric arcs.

The indices are based on the elements of the folds shown in right section (Fig. 15), and are referred to its vertical and horizontal ordinates. The ordinates are drawn through the fold centres. Fold centres O^A and O^S of the anticline and syncline are separated by the horizontal distance B, the width of the fold in right section. The radii of folding of a stratum* are given by R_a and R_s for the anticline and syncline respectively. The angle θ lies between vertical ordinates drawn through the fold centres of the anticline and syncline, and the radius normal to the stationary tangent at the point of contraflexure T of the folded stratum. The fold amplitude A is measured along the vertical ordinates of the right section. The anticline is contained in the arc defined by the angle θ, and its amplitude is given by A_a. The anticlinal amplitude is the vertical interval of the folded stratum, between its intersection with the vertical ordinate of the anticline and T the point of contraflexure, in section.

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* A stratum is considered to have negligible thickness for the purpose of this discussion.

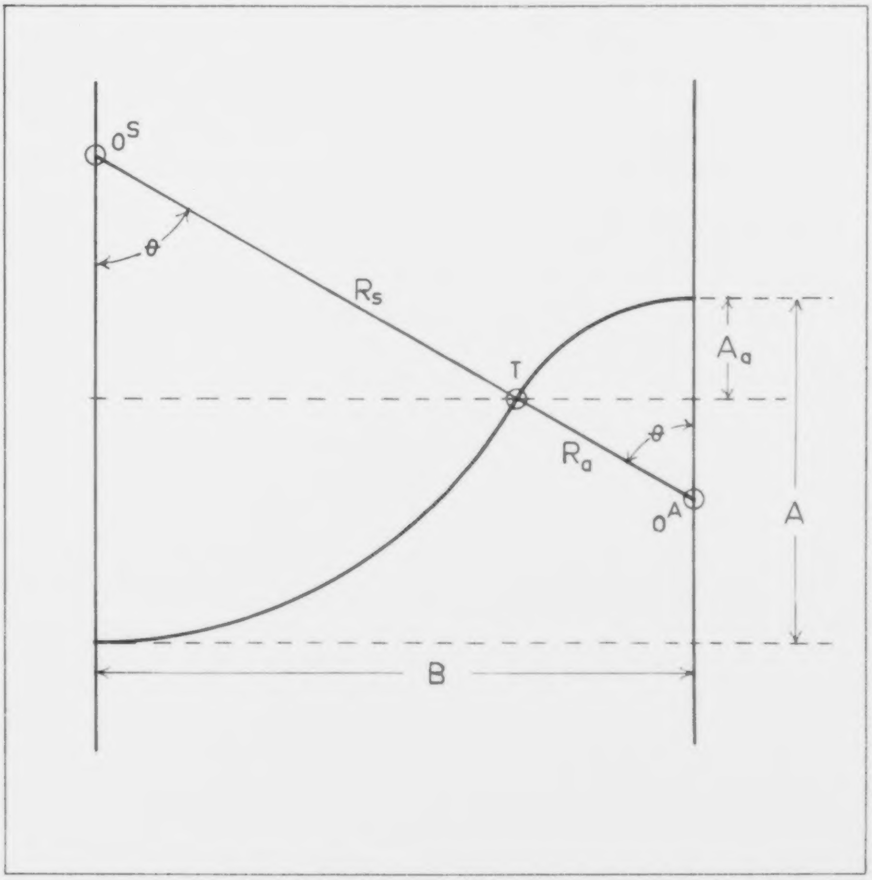

Figure 15. Fold and anticlinal amplitudes in concentric folding: symmetrical fold. Explanation in text.

The index of fold amplitude is the ratio A/B. The index A/B is constant for any stratum in a given fold (by construction). In any series of folds for which θ is the same, and R_a/R_s varies, the index A/B remains constant. The index A/B depends on θ. In Figure 15,

$$\begin{aligned} A &= (R_a - R_a \cos\theta) + (R_s - R_s \cos\theta) \\ &= (R_a + R_s)\ (1 - \cos\theta) \\ B &= R_a \sin\theta + R_s \sin\theta \\ &= (R_a + R_s) \sin\theta \end{aligned}$$

$$A/B = \frac{1 - \cos\theta}{\sin\theta}$$

The amplitude of the anticlinal fold $A_a = R_a \, (1 - \cos\theta)$. The index of anticlinal amplitude is $(1 - \cos\theta)$ for any stratum in a fold, and is independent of R_a.

Different folds and their amplitudes can be compared by the indices A/B and $(1 - \cos\theta)$. The indices are independent of R_a and R_s, and obtained from any stratum in the folds (that is, concentric folds, and therefore strata between fold centres of anticlines and synclines).

Practical Use.—The indices A/B and $(1 - \cos\theta)$ apply to right sections drawn perpendicular to the axis. Their values are obtained in such sections for cylindrical folds, with or without plunge, and for non-cylindrical folds. The indices then become arbitrary measures of the fold form. The amplitude of folds vary along their length and decrease to their terminations. Comparisons of fold and anticlinal amplitudes by the indices A/B and $(1 - \cos\theta)$ require sets of serial sections, or the choice of typical sections. Usually the median right section of the fold is suitable. Concentric folds in nature depart from the simplified condition of the discussion, by the occurrence of subsidiary structures in the limbs and imperfections of concentricity. It is considered that these departures do not affect the reliability of the indices of amplitude of folds and anticlines. Smoothed curves are drawn through subsidiary folds and faults of minor order. Indices of fold and anticlinal amplitude for concentric folds which depart from simple geometric concentricity become empiric ratios but remain effective. The relation of fold and anticlinal amplitudes to θ is borne out by field observations, anticlines of low and high amplitude forms having low and high values of θ respectively.

Symmetric Folds.—The index of fold amplitude and the index of anticlinal amplitude are shown by the symbols A/B and $(1 - \cos\theta)$ according to the foregoing discussion.

Asymmetric Folds.—The folds are referred to horizontal and vertical ordinates in right section. The fold amplitude A is given by the vertical interval of a folded stratum measured between the crest of the anticline and the point of maximum depression in the syncline, in section. The same points of reference define the horizontal width of the fold. The anticlinal amplitude is measured between the crest of the anticline and the point of contraflexure for any stratum (Fig. 16).

The asymmetric fold is still referred to the ideal condition of concentric arcs as the theoretical model, and the indices of fold and anticlinal amplitude derived as for symmetric folds. It is considered that in practice the asymmetry of a fold does not appreciably affect the accuracy or validity of these indices.†

Both indices of fold and anticlinal amplitude in asymmetric structures differ for the halves of the fold and anticline separated by the vertical ordinate at the anticlinal crest (Fig. 16). The indices of the two fold halves are determined as A/B * (As S), and $(1 - \cos\theta)$ * (As S) for the steep-dipping half of the fold; A/B * (As L), and $(1 - \cos\theta)$ * (As L) for the less steep-dipping half of the fold.

Faulting.—Asymmetric anticlines are commonly faulted across the steep limb. in the case of faulted sectors, the index of anticlinal amplitude $(1 - \cos\theta)$ is best disregarded, for the principle of concentricity does not apply to faults.

The index of fold amplitudes A/B * (As) can be maintained, and used in two ways. In the first, it is used directly from field observations. The index is taken from the beds displaced across the fault. It then accounts for displacement of strata in folding and faulting combined. The index is designated the *index of amplitude of fold and fault movement*, and shown as A/B * (As F). In the second way, the structure is reconstructed to an unfaulted condition, as shown in Figure 16. The index A/B becomes the *reconstructed index of fold amplitude* and denoted by A/B * (As R). Where displacements on the fault vary and field observations are sufficient, the indices A/B * (As F) and A/B * (As R) are determined for a range of values.

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† The axial planes of asymmetric folds seem attractive as bases for "natural" ordinates. However, they are located arbitrarily, and usually their subsurface traces must be assumed. The dip of any axial plane may vary; axial planes of an anticline and its adicent syncline may have different dips. Axial planes are impractical and uncertain reference ordinates.

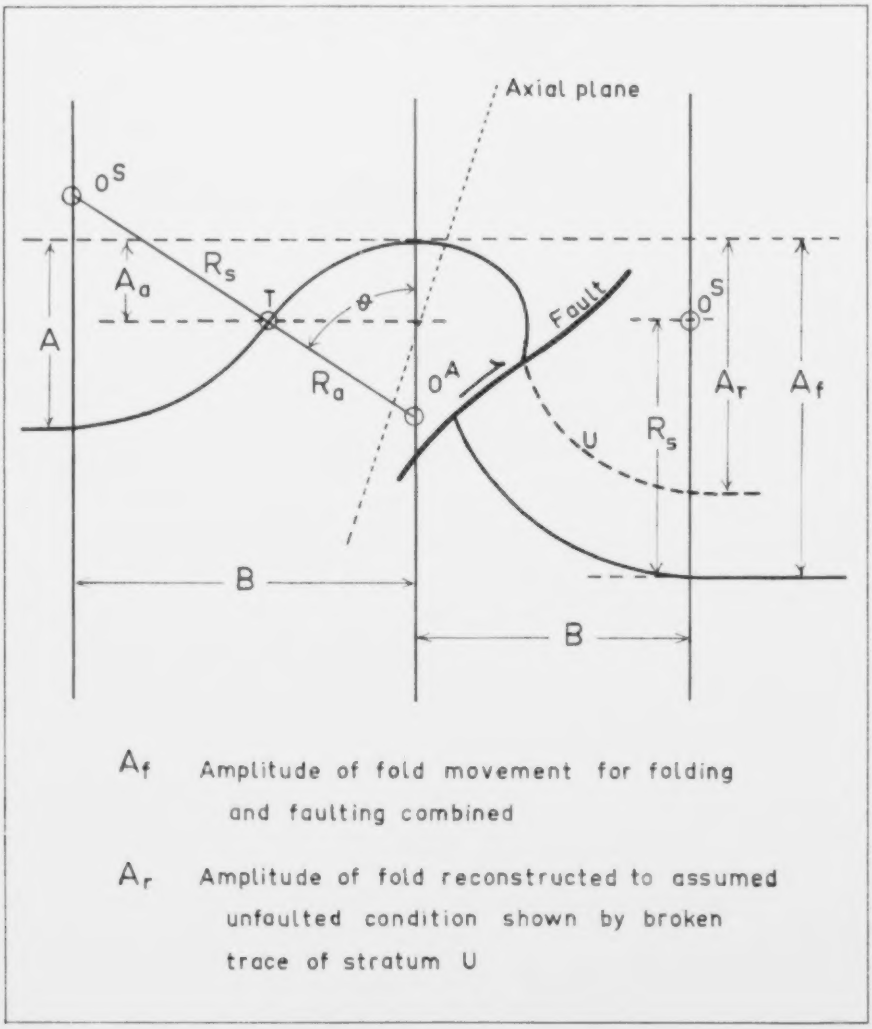
Axial plane
O^S
R_s
A_a
T
θ
A
R_a
O^A
Fault
U
O^S
A_r
A_f
R_s
B
B
A_f Amplitude of fold movement for folding and faulting combined
A_r Amplitude of fold reconstructed to assumed unfaulted condition shown by broken trace of stratum U

Figure 16. Fold and anticlinal amplitudes in concentric folding: asymmetrical faulted fold. Explanation in text.

CONCENTRIC ANTICLINES OF HIGH AND LOW FOLD AMPLITUDES

A summary distinction of these fold types in general, by use of selected constants of the indices A/B and $(1 - \cos \theta)$, may or may not be possible. The two types of anticlines are readily separated in the Pine, Moberly, and Pine River areas, by the values of the indices taken from representative or median sections. These values are listed in Table II. It is found that anticlines of high fold amplitudes have $A/B > 0.25$; $(1 - \cos\theta) > 0.25$; and $\theta > 40°$. These values are considerably higher for anticlines of the Inner Foothills. Anticlines of low fold amplitude have $A/B < 0.1$; $(1 - \cos\theta) < 0.025$; and $\theta < 12°$.

Table II.—Indices of Fold and Anticlinal Amplitude for Concentric Anticlines in the Foothills of the Peace, Moberly, and Pine River Areas

Anticlines	A/B	$(1 - \cos\theta)$
1	0.590	0.620
2A	0.985	1.000
2B	0.780	0.4554
3	0.905	0.5548
4	1.000	0.5550
5	0.295	0.3026 to 0.4264
6	0.076	0.0097 to 0.0055
7	0.0673	0.0030
8	0.0192	0.0024
9	0.0883	0.0152

High Amplitude Folds

1. Big Boulder Creek anticline, Figure 3. Indices (AsL).
2A. Cairns anticline, Figure 3. Indices (AsS).
2B. Cairns anticline, Figure 3. Indices (AsL).
3. Silver Sands anticline, Figure 3. Indices (AsL).
4. Monteith anticline, Figure 14. Indices (AsL).
5. Butler Ridge anticline: section AB of Beach and Spivak (1944). Indices (AsL).

Low Amplitude Folds

6. Commotion anticline, Figure 3. Indices (AsL).
7. Chetwynd anticline, Figure 3, section E.N.E. at Chetwynd (MSS., writer, 1956). Indices (AsL).
8. Hudson Hope anticline, section E.N.E. through Lower part of the Peace River Canyon (MSS., writer, 1959). Indices (AsL).
9. Hulcross anticline: Median section, after Wickenden and Shaw (1943). Indices (AsL).

CONCENTRIC ANTICLINES OF HIGH FOLD AMPLITUDE AND THEIR DEVELOPMENT

The concentric folding of an anticline depends on an equilibrium of growth provided by (1) arcuate bending of strata about a fold axis; (2) the rise of the fold centre; (3) the replacement of the arcuate bending by non-concentric deformation from below. Thick competent formations bend and transmit the fold compression (Willis, 1923, pp. 78–94; Nevin, 1942, p. 50). Non-concentric deformation occurs underneath, within competent or incompetent formations.

As the fold is compressed, alternative developments of folding are possible (Fig. 17). In the first, concentric folding becomes replaced by a complex of folding in non-parallel and parallel forms, shear and faulting, or by simple cuspate folding (Fig. 17-2a). Cuspate folding is treated in following discussions.

In the second development (Fig. 17-2b), the folding tends to a disequilibrium of the growth condition already noted. In this development, the relative movement of the fold centre decreases and it becomes fixed on one stratigraphic plane, usually about the base of a relatively thick competent sequence. Concentric folding is completed about this stationary fold centre, as the fold limbs are closed; the anticlinal fold is pushed up vertically. The disequilibrium condition is brought about by (1) the lack of space in the core of the fold as it narrows (this has been pointed out by de Sitter 1956, p. 199), and (2) thick competent formations which permit bending, but tend to resist non-concentric deformation and the rise of the fold centre.

Fold centres of some major anticlines are limited by competent formations, for example, the Monteith Formation in the Portage Mountain anticline (Fig. 12) and the Triassic Grey Beds in the Silver Sands and Cairns anticlines (Fig. 2). This field association is not invariable, for non-concentric folding and other deformation of competent beds and formations occurred.

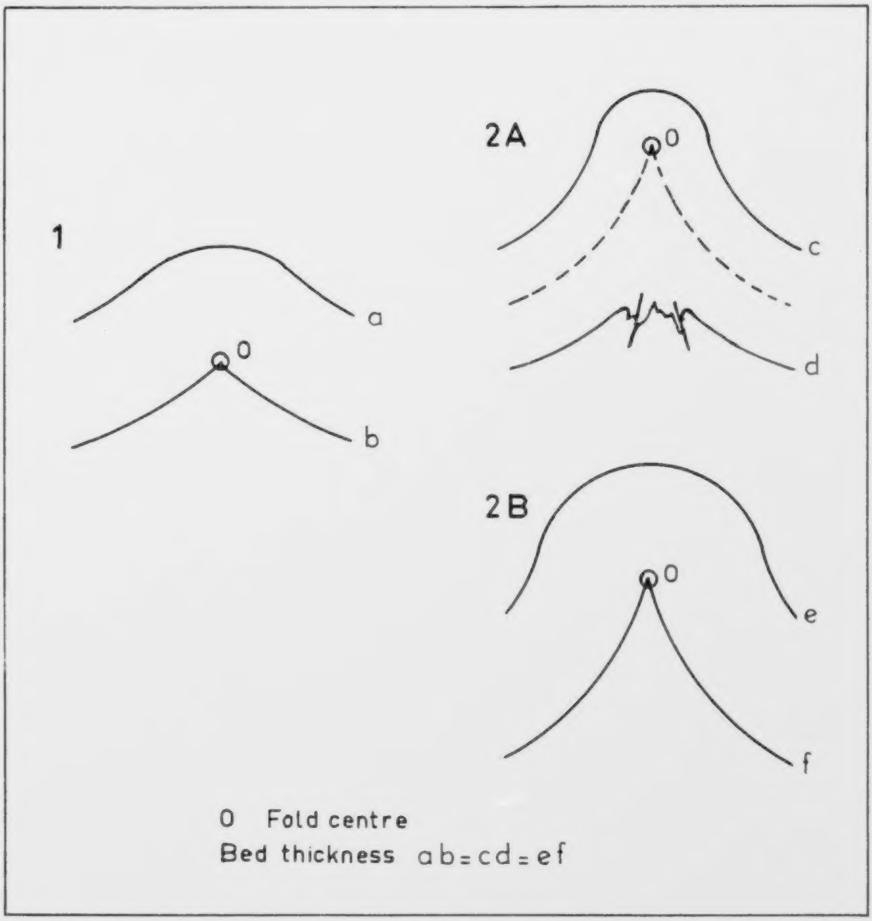

Figure 17. Alternative developments, 2A and 2B, of a concentric fold from an initial fold stage 1. In 2A the fold centre O moves upwards into higher beds as the fold is compressed. In 2B the upward movement of the fold centre is limited by the competent beds of the interval *ef*.

Both developments of folding at some advanced stage can produce concentric anticlines of high fold amplitude. The extreme forms of high amplitude folds, the subcircular and box forms, are considered to result from the second development indicated. In this, closing of the fold limbs under compression and the vertical movement of the fold, together with restriction of space in core of the fold, bring about faulting in later stages. Thrust faulting with attenuation and distortion of

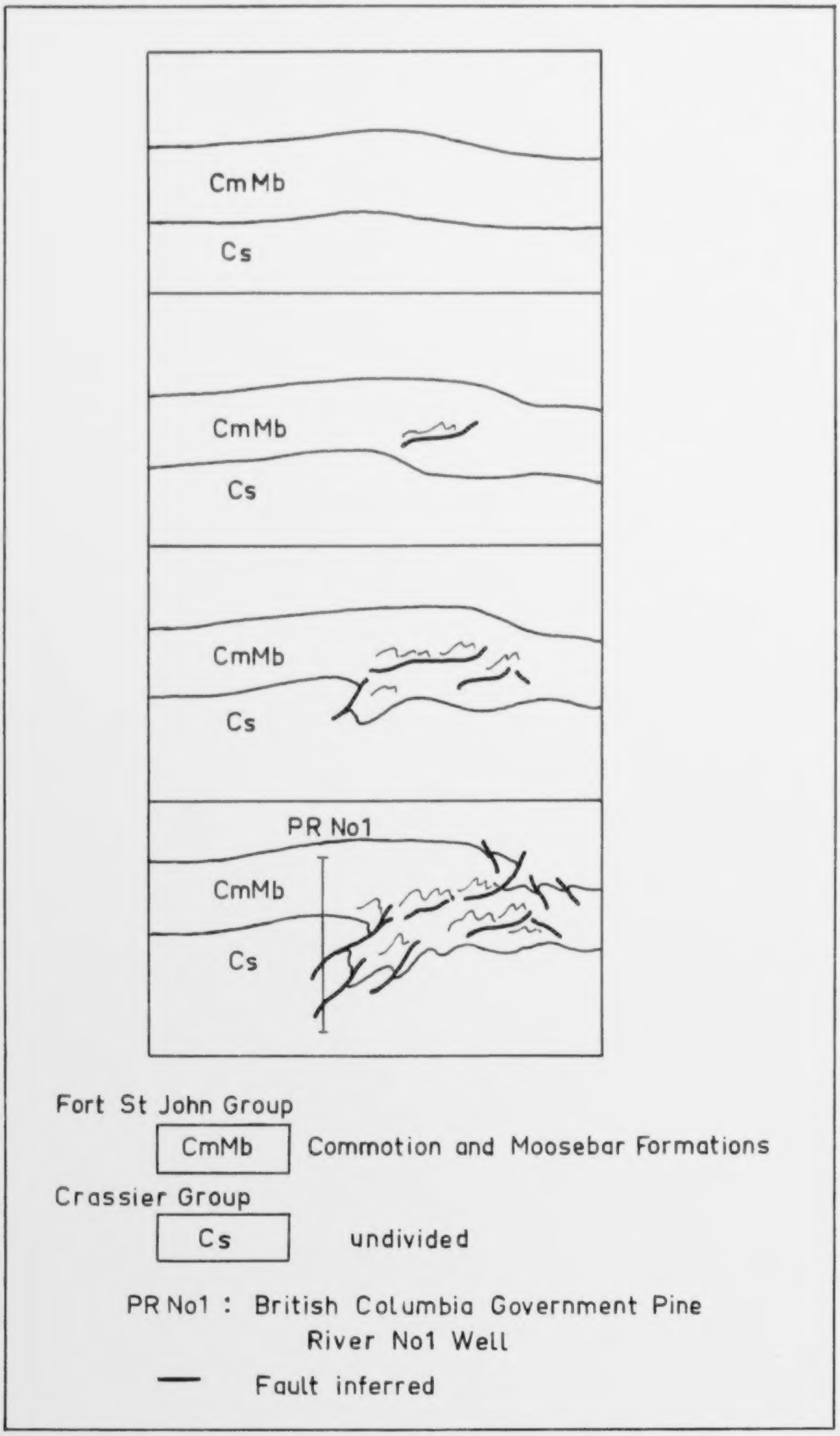
CmMb
Cs
CmMb
Cs
CmMb
Cs
PR No1
CmMb
Cs
Fort St John Group
CmMb Commotion and Moosebar Formations
Crassier Group
Cs undivided
PR No1 : British Columbia Government Pine River No1 Well
Fault inferred

Figure 18. The development of the Commotion anticline (theoretical).

beds in the steep limb obtains in asymmetric anticlines. This is the common rule in the Foothills and can be seen in many examples already given. Diapiric movement of the axial part of the anticline is possible, as noted by de Sitter (1956, p. 199). The Monteith anticline shows an approach to this condition (Fig. 14).

CONCENTRIC ANTICLINES OF LOW FOLD AMPLITUDE AND THEIR DEVELOPMENT

Concentric anticlines of low amplitude are less easily studied, for their fold centres and cores lie concealed in the subsurface. A full understanding of these structures requires the observation of drilled sections. The Commotion anticline is the only one sufficiently known. Others resemble it in their surface form, but some differ slightly in being compound and containing accessory folds and faults of minor order.

The Commotion structure is treated in Chapter IV. Its mode of folding is indicated briefly here. The upper part of the structure underwent movement and bending independently, or partly so, of structures in the lower part. The two parts of different structure were separated by a thrust, or complex of multiple small folds, faults, and shears, with partial décollements. This structural break occurred within incompetent shales and mudstones, the folding of the overlying part in alternate competent and incompetent formations. The break between the two parts and their different folding originated in early stages of the fold compression. A sequence illustrating the fold development is given in Figure 18.

All other anticlines of low amplitude in the Outer Foothills occupy the same tectonic environment as the Commotion anticline. They were also formed in alternating competent sandstones and incompetent shales of the Fort St. John Group and the Dunvegan Formation. Their structures at depth may resemble the Commotion structure but remain open to proof.

NON-CONCENTRIC PARALLEL FOLDS

Much of the folding is non-concentric. Many angular folds are recognized with the following properties:—

(1) A sharp angular bend at the axis—which is maintained throughout the fold or within an interval of a fold.

(2) Parallelism of beds.

(3) Straight limbs with uniform dips, or limbs with moderate, slight, or negligible curvature in the limbs.

These are the common properties of angular folds. These folds also have the following individual features, as noted in field examples:—

(1) A fault close to the axis. The fault dips parallel or nearly parallel to the axial plane. Examples include minor folds in the Coyote Creek syncline, and a small syncline east of the Monteith anticline—this fault is inferred from the displacement of beds across the axial part of the fold (Fig. 14).

(2) A thrust fault cutting the steeper limb close to the axial plane. Thrust faulting can be confirmed in one example, the Bickford anticline (Figs. 2 and 19), and can be inferred from the Mount Gething-Stott Creek anticline (Fig. 13, from the mapping of Beach and Spivak, 1944). Both examples are major structures.

(3) Planes of separation and movement along the bedding, that is, "bedding plane thrusts." This condition is shown in a small fold in beds of the Crassier Group (Fig. 20) and in a fold of the Grey Beds (Fig. 21).

(4) The composition with an overlying concentric fold in one structure. This condition occurs in the Mount Gething-Stott Creek anticline, according to

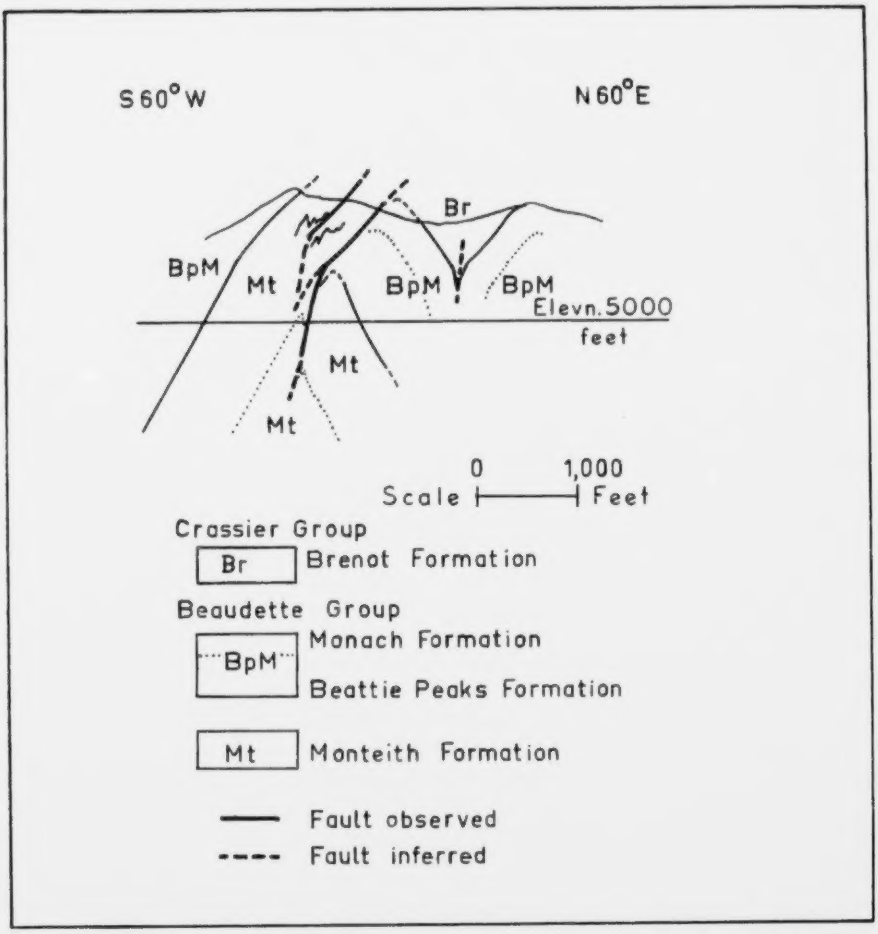
S 60° W
N 60° E
Br
BpM
Mt
BpM
BpM
Elevn. 5000 feet
Mt
Mt
0
1,000
Scale
Feet
Crassier Group
Br
Brenot Formation
Beaudette Group
BpM
Monach Formation
Beattie Peaks Formation
Mt
Monteith Formation
Fault observed
Fault inferred

Figure 19. Section of the Bickford anticline, 2¼ miles north of the summit of Mount Bickford.

Beach and Spivak (1944, p. 12). The core contains " strata tightly compacted in a narrow zone of intense deformation." It is an angular, probably a cuspate fold, in section, and from the mapping. The overlying part of the fold has a " broadly arched crest." This folding was concentric. The whole fold form is composite. The anticline is cut by two faults in the steep limb (Fig. 13).

(5) Angular folding without any appreciable faulting along the fold axis—discounting small scale deformation restricted to very incompetent beds such as coal seams. Serial sections, taken from the results of drilling and stripping, and reported by McKechnie (1955), provide an example (Fig. 22). Other folds of this type are the Folded Hill structure mapped by McLearn (*see* in McLearn and Kindle, 1950, Fig. 8), and folds in Crassier beds at Beaudette Creek.

(6) A deflection of strata between the main part of the limb and the axis of the fold. This deflection may or may not accompany a fault. Usually,

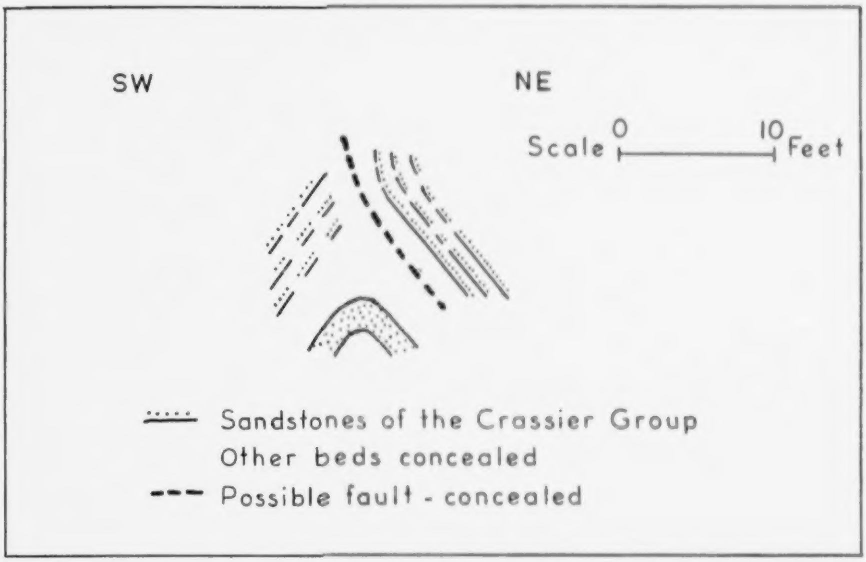

Figure 20. Diagram of a minor fold, east flank of the Bickford anticline, on the north wall of the Pine Valley, elevation 3,650 feet.

the beds increase in dip toward the fold axis, as in folds at Noman Creek (Fig. 22). The west limbs of the Bickford and Fisher anticlines show a decrease of dip near the axis.

Angular folds and their classes have been already defined. A distinction between lambdate and cuspate folds is not always practicable in the field. It depends on exposure and the vertical range of observation. Where this distinction is uncertain, the general term angular folds must be applied.

Lambdate folds have straight limbs and a " sharp anticlinal hinge " which is continuous (compare the description of accordion folds by de Sitter, 1956, p. 231). A few accordion folds of this class are recognized. They are subsidiary structures and include minor folds of the coal measure in the Coyote Creek syncline, and folds of the Beattie Peaks Formation in the west flank of the Bickford anticline. Some accordion folds were associated with thrust faulting as "dragfolds," for example, those in the footwall of the Pyramis thrust in the hills north of the Pine River, and those in quartzites of the Monteith Formation about the main thrust fault of the Bickford anticline, north of the Pine River (Fig. 19). One fold, the syncline east of Mount Monteith (Fig. 14), may be referred to the lambdate class.

The following criteria serve to identify cuspate folds in the field: (1) Curvature of the limbs; (2) composition with overlying concentric folds in anticlines, and underlying concentric folds in synclines; (3) a change from cuspate form to concentric form along the same axis of folding. Cuspate folds are closely associated with concentric folds in the Foothills of the Peace and Pine River areas. Such association can be expected to be general. Deflection of the limbs near the fold axis appears to be characteristic of some cuspate folds. The Mount Gething-Stott Creek anticline is a distinct example according to the above criteria (Fig. 13). Folds of this class in the Pine River area include the Noman anticline (Fig. 22); some minor folds 1.5 miles upstream at Narod Creek, and other small folds between Fisher and Crassier Creeks; and minor folds at Beaudette Creek.

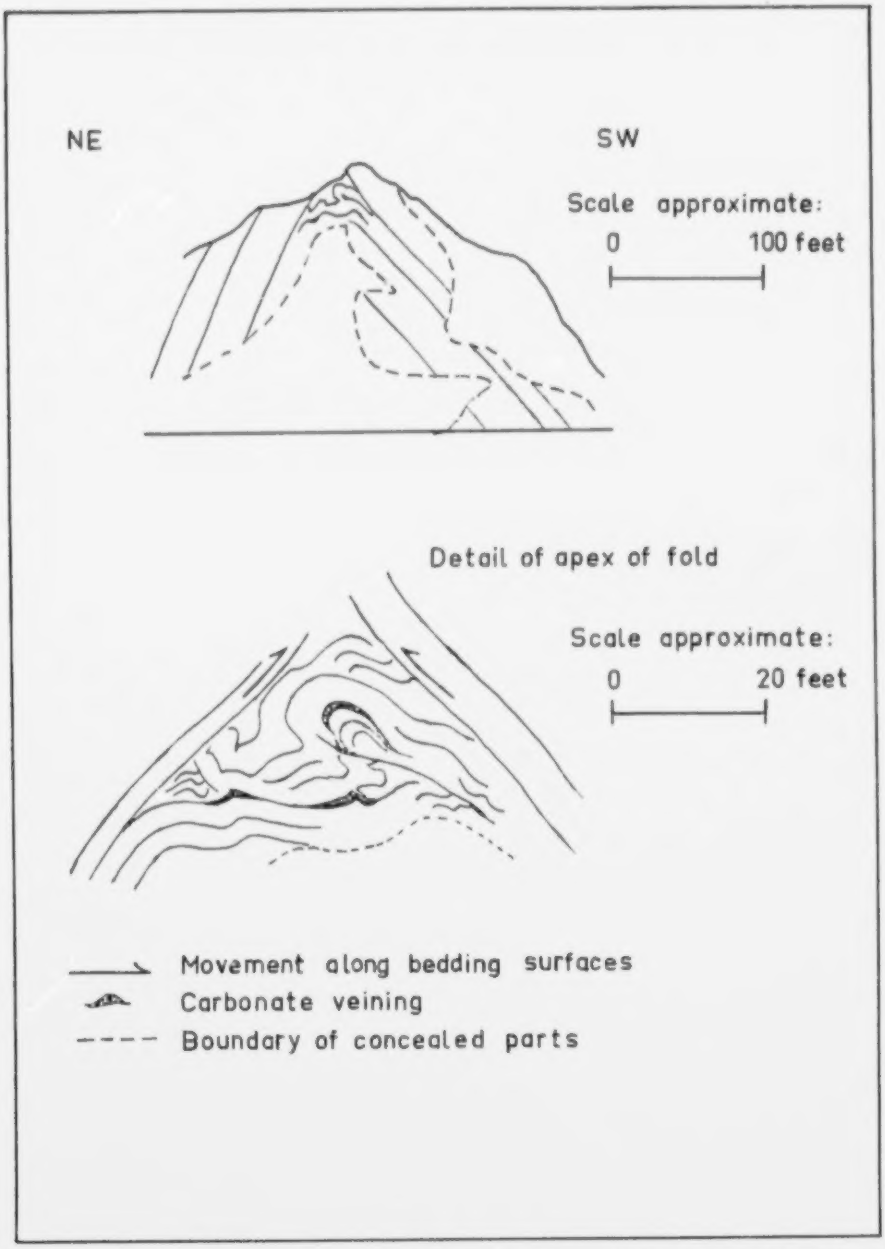
NE
SW
Scale approximate:
0
100 feet
Detail of apex of fold
Scale approximate:
0
20 feet
Movement along bedding surfaces
Carbonate veining
Boundary of concealed parts

Figure 21. Angular fold in the Grey Beds, near the railway bridge at Mountain Creek, Pine Valley.

Both lambdate and cuspate folds may have axial faults. In lambdate folds, movement on the axial fault may be normal, reverse dip slip, or oblique slip. In cuspate folds, such faults are thrusts according to observations in the Pine Valley.

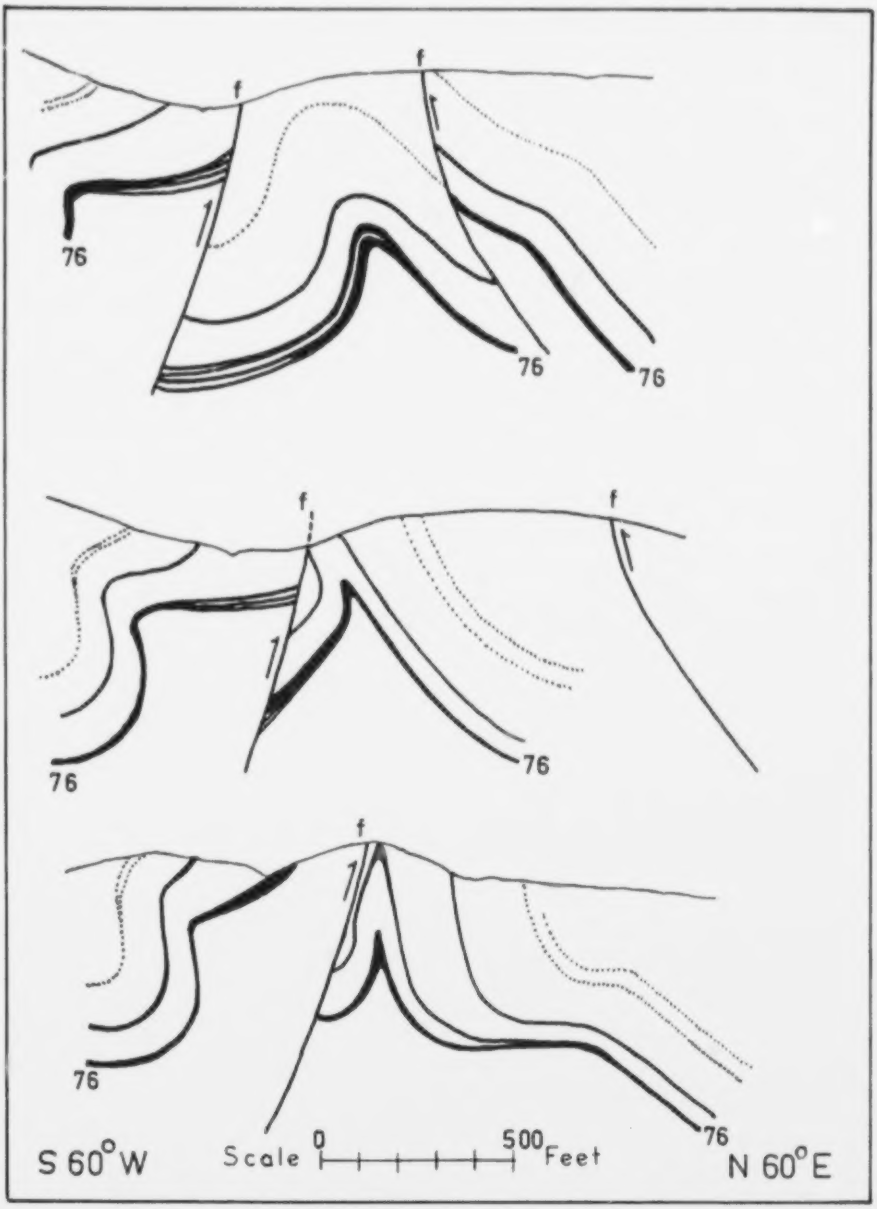

Figure 22. Serial sections of the Noman Creek structure, Pine Valley, after McKechnie (1955). Sections are 1,000 and 1,400 feet apart. f: Fault. 76: Coal seam No. 76.

This association may be local. It is uncertain whether it applies to cuspate folds in general.

Some structures are better described as angular folds. The anticline shown in Plate IV is an example. Its west limb is virtually straight for most of the observed section; the east limb is curved. The anticline at Mountain Creek appears lambdate in its gross form, but the apex is disharmonically folded (Fig. 21). The Fisher anticline (= Pine River anticline of Spieker, 1922, p. 22) is also described as an angular fold. It lies between Crassier and Fisher Creeks, 3 miles north of the Pine River. It is a major structure, 2.5 miles wide. At the apex, beds of the Monach Formation close in a sharp bend which can be crossed over in one pace. Spieker commented on the " sharpness " of this structure. The Bickford anticline, classified as angular, may represent a cuspate fold—its upper structure above the 5,000 foot level is considered to mark the core of a concentric anticline which has been eroded (Figs. 2 and 19).

In other folds, non-concentric fold forms are discontinuous and only partly developed in the structure. The structures retain parts of concentric folding; the other parts, which are cuspate folds, have limbs with slight curvature (dips increasing to the fold axis), and commonly a fault break near the axis. These structures may belong to the composite folds of Hills (1953, p. 94). The examples at hand are obscure for lack of continuous exposure. A group of minor folds in the Crassier beds east of Mount Pyramis are angular in the upper levels at 4,500 feet elevation. At lower levels they are more simple, rounded, and broad. The form of the Crassier anticline is concentric in exposures at river level. It is angular, three-quarters of a mile northwest, and 1,500 feet higher in elevation. Little can be seen of the ground between the two sections. The fold maintains the same asymmetry and axial plane, which dips northeast. Some of the folding in coal measures in the east flank of Bickford anticline, on the north face of the Pine Valley, is probably composite.

LAMBDATE AND CUSPATE FOLDS: MECHANISMS

The mechanisms of cuspate and lambdate folding are partly alike. They both fold by a sharp angular bend across the axial plane. The folding requires shear and slip along the bedding. Faulting often develops along the axial plane in advanced stages of folding; minor faulting may occur in the limbs.

Lambdate folds are considered to develop uniformly after they are initiated within a mass of strata. They first form by the sharp bend at the axis. This bending occurs at all levels of strata at the same time, and each stratum maintains its own bend at the axial plane. The limbs are closed by compression and the fold is dilated vertically. The relative trajectories of points in folding can be shown, depending on alternative assumptions. In the first, a reference bed B of constant length is moved with continuous contact over a datum plane (Fig. 23-A). In the second, a reference bed B is folded about a datum point O (Fig. 23-B). In each assumption the trajectories show that one component of the movement is directed inwards to the axial plane.

Cuspate folds develop in the cores of concentric folds undergoing compression. For an anticline the upper beds are bent concentrically above a fold centre. The radius of folding of any bed decreases in compression, and the fold centre moves upward in stratigraphic level. At the fold centre the concentric folding of any bed becomes replaced by angular bending (its radius of folding being infinitely small). This replacement is progressive, bed for bed, as folding continues, and the fold centre rises.

A geometric model* (Fig. 24) shows the sequence of the folding. Relative trajectories of folding are shown. Beds of the angular fold below the fold centres

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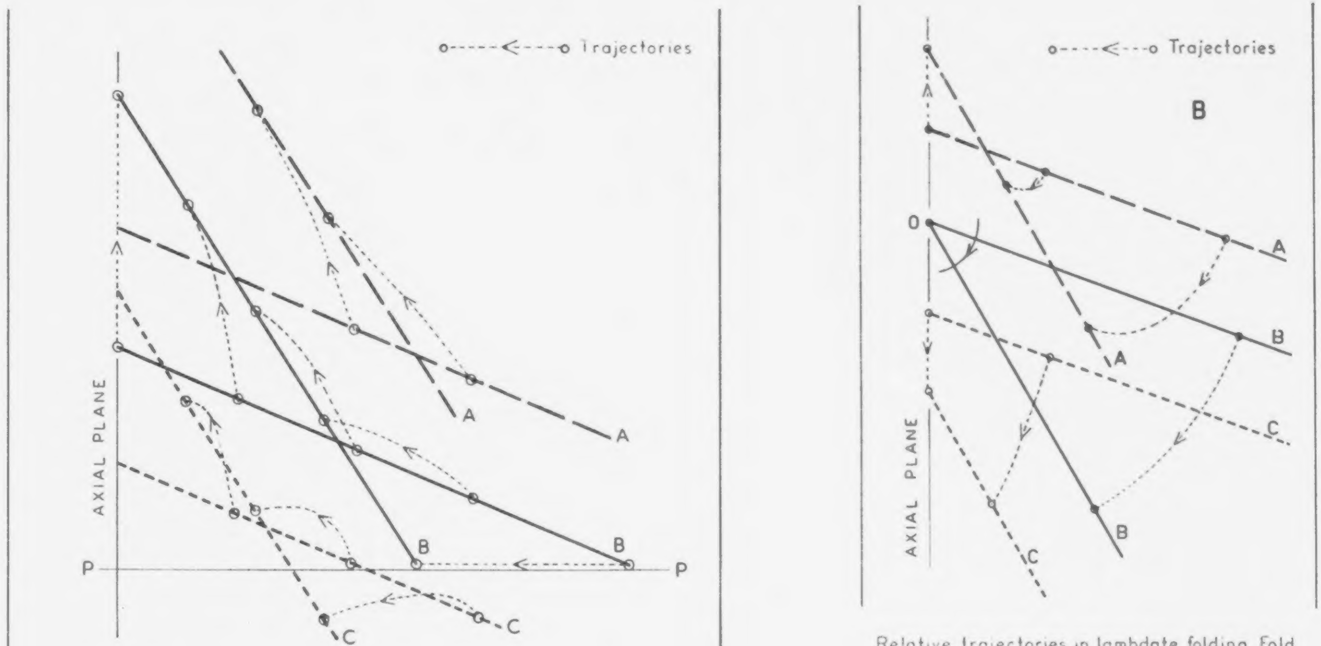
* This model is based on the following assumptions: Constant bed thickness and equal length of all beds is maintained in the folding; the fold is represented by concentric arcs; the sum of the radii of synclinal and anticlinal folds is constant; successive fold centres of the syncline lie in a plane parallel to the datum plane of the fold. Another model is possible, for which the radii of synclinal and anticlinal folding vary in constant proportion, and for which the successive fold centres of the syncline are not co-planar; this model shows the same results as the one illustrated in Figure 24.


Trajectories
AXIAL PLANE
P
P
A
A
B
B
C
C
Trajectories
B
0
A
A
B
B
C
C
AXIAL PLANE

Relative trajectories in lambdate folding. Fold movement is by sliding of bed **B** along a datum plane **P**. Trajectories are shown for points on the parallel beds, **A**, **B**, **C**, in the section of the semi-fold.

Relative trajectories in lambdate folding. Fold movement is by rotation of bed **B** about a datum point **O**. Trajectories are shown for points on the parallel beds, **A**, **B**, **C**, in the section of the semi-fold.

Figure 23. Relative trajectories in lambdate folding.

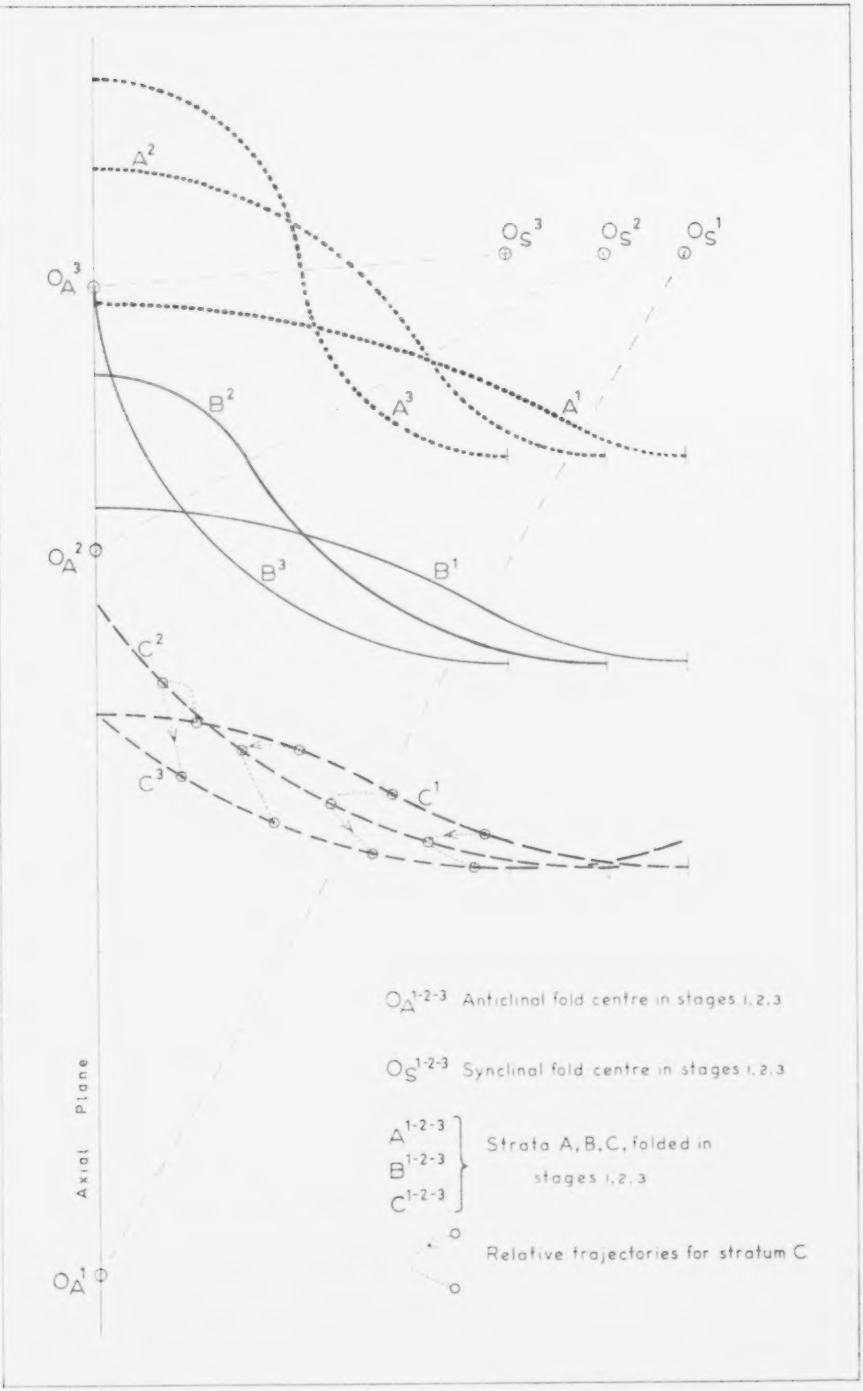

A^2
O_S^3
O_S^2
O_S^1
O_A^3
A^3
A^1
B^2
O_A^2
B^3
B^1
C^2
C^3
C^1
O_A^{1-2-3} Anticlinal fold centre in stages 1, 2, 3
O_S^{1-2-3} Synclinal fold centre in stages 1, 2, 3
A^{1-2-3}
B^{1-2-3}
C^{1-2-3}
Strata A, B, C, folded in stages 1, 2, 3
Relative trajectories for stratum C
Axial Plane
O_A^1

Figure 24. A model for the development of cuspate folding, shown for the beds A, B, C in a semi-fold.

are displaced down and out. In the lower part the beds are displaced beyond the vertical boundaries of the fold model, that is, the boundaries constructed for the related overlying concentric fold.

Compression of cuspate folds is obtainable by a contraflexure of beds, or by reverse faulting in the limbs. Either mechanism allows the constraint of the fold within approximately the same span as the related concentric fold above. Alternatively, in much of the cuspate folding, a proportion of the mass of strata in the core of the fold is excluded, and redistributed and accommodated in adjacent structures—concentric, disharmonic, and angular folds, and fault displacements.

DEVELOPMENTS OF CUSPATE FOLDS

The growth and possible developments of cuspate folds in compression have been deduced from field observations of comparative structures representing different stages of folding, together with reconstructions of fold movement in geometric models. These reconstructions have been concerned with the distribution of a mass of strata in the folding, such that thickness and length of beds in cross-sections remain constant. The results of a theoretical treatment based on such models become inconclusive. Many conditions exist for the models, depending on the axial planes and the boundaries specified in the folding.

The boundaries in which beds of cuspate folds were contained are indeterminate for field examples in general. One boundary of the semi-fold (between anticlinal and synclinal axes) of cuspate form may be presumed as the plane containing the axes of angular bending in anticlines. The other boundary, or reference plane, in the semi-fold cannot be presumed as the axial plane of a syncline which is concentrically folded in a curve because there is no known limit to the length of the bed in such cases.

Vertical parallel plane boundaries of beds in the folding of a single structure, as in simple test boxes, are not entirely representative of natural conditions. Involutes, or concentric arcs representing beds folded in cuspate folds, have unequal intercepts between vertical and inclined parallel plane boundaries, as shown in sections. On the other hand, accordion folds can have beds of equal length between parallel plane boundaries.

In geometric models of cuspate folds, parallelism and equal lengths of beds cannot be maintained within limits of parallel plane boundaries, nor otherwise in the compression of the models by closing the distance between the boundaries. This means that the fold form is unstable in the arbitrary boundary conditions of the test—that it tends to become modified or destroyed by faulting, other folding or distortion, to accommodate the mass of strata under compression, within the boundaries denoted.

In cuspate anticlines, the radii of folded beds increase at depth. Then intercepts of circular arcs or involutes representing the beds between parallel plane boundaries approach equal lengths. With corresponding decrease in curvature of the beds, the forms of cuspate folds approach those of accordion folds; distinctions between cuspate and lambdate forms become less apparent. The angle between the fold limbs across the axial plane is constant in accordion folds, but varies in cuspate folds, decreasing downwards in anticlines.

No useful purpose is served by providing geometric models of developments of cuspate folds with reference to arbitrary boundaries. Instead, a summary account of the possible developments of these folds may be given, with comparisons from field evidence where this is appropriate and available.

(1) The cuspate fold undergoes compression by a decrease in the radius of folding of beds.

(2) The cuspate fold undergoes compression by the mechanisms indicated for lambdate folding.

Both developments increase the acuteness of the fold in its upper part and tend to break open the fold about its axial plane. There is no direct evidence for the developments (1) and (2) above. However, distortions of some cuspate and angular folds about the axis, faults which split the fold along the axial plane, and extreme acuteness of folding have been noted in some examples (Figs. 19, 20, 21, and 22).

(3) Cuspate folds alternate with concentric folds vertically in unit structure.

(4) Beds in the limbs of the cuspate fold undergo contraflexure, possibly in conjunction with either of the mechanisms indicated under (1) and (2). The contraflexure permits the cuspate fold to be contained in approximately the same width at all levels.

Evidence for the types and developments of cuspate folds in the possibilities (3) and (4) is incomplete, as indicated previously for the folds east of the Pyramis thrust and the Crassier anticline.

(5) The cuspate fold is modified by thrust faulting along the axial plane. Thrust faulting originates at depth in the fold. It extends upwards as the structure is compressed, while angular folding replaces the concentric fold above. Thrust faulting in combination with folding provides for the contraction of the structure. The thrust finally joins superficial faults of the overlying concentric anticline. Folding may cease at this stage, later compression of the structure being resolved by fault movement.

In one major structure, cuspate folding was modified by thrust faulting about the axial plane, in the Mount Gething-Stott Creek anticline (Fig. 13); the Bickford anticline may provide another example (Fig. 3).

FOLDING: GENERAL REMARKS

Some relationships of concentric and similar folding were considered by Ez (1959) and Govsky (1959). Ez believed " that concentric folding may be accompanied by non-concentric folds into which they grade vertically," and that " both types of folds are merely the result of arresting the process of folding at different stages." He obtained, in experiment, a sequence of concentric folds passing to similar folds with continued compression. Ez also noted the association of similar and concentric folds in the same strata in one area. Govsky examined some physical factors affecting concentric and similar folding, and reported them from experiments. The types of similar folds are not classified. Govsky observed " composite folds," with combined similar and concentric fold characters, in the experiments.

The physical factors, stress in shearing and compression, rate of stressing, which obtained in concentric folding and the types of angular parallel folding discussed, are indeterminate from geological structures. Few generalities about geological factors affecting the development of angular folds can be derived from field relations, and must be expressed tentatively.

Angular folds occur in most strata of the Foothills below the Fort St. John Group. Such folds, on a small scale, are numerous in thin-bedded shales and sandstones and coal measures. The incidence of angular folding in these strata indicates the importance of bedding slip in fold movement. Otherwise, or at some stage in compression, faulting provided movement within a fold, perhaps more so in thick competent formations.

The few examples classed as accordion folds are minor structures (mostly " dragfolds "). Most of them appear to adjoin thrust faults. As far as observed (writer, Pine Valley), the accordion folds are without definite or regular cleavage, which is extensive throughout the fold. A break or fracture of the fold along the axial plane is characteristic. The same remarks apply to other folds referred to the lambdate class.

No regular or extensive cleavage appears in cuspate folds, nor angular folds of indeterminate class. The internal deformation of these folds includes faulting along the axial plane, thrust faulting cutting limbs at low angles, and slip and translation between beds. A restricted cleavage in weak beds, such as coals, may be implied by the distorted form of some cuspate folds about the axis. Exposures which may demonstrate this condition conclusivly have not been found; coal seams are severely weathered and rotted in the examples noted, or concealed. Exclusion and displacement of some of the mass of strata from the core of the cuspate fold is considered, a movement which tends to reduce the internal strain of the fold and inhibit cleavage.

Elsewhere, small developments of cleavage are sparingly distributed, in shales of the Fernie Group, in Triassic beds, and in the Fort St. John Group. These few occurrences lie in parts of folds, jammed in the footwalls of thrusts. Cleavage is also found in shales between thicker competent beds, sandstones and siltstones, in both concentric and angular folds.

VARIATIONS OF FOLD FORMS AT DEPTH

The foregoing discussions refer to variations of fold forms with depth. The subject may be reviewed here as an aid to mapping and petroleum exploration.

Structures underlying concentric anticlines may be either cuspate folds in several possible developments, perhaps some forms of similar folding, or complexes of multiple faulting and folding. In usual concepts, concentric folding is thought to require décollements at depth, either locally below anticlines or on extensive detachment planes. The concentric folds in the Foothills are known to be superficial, but they are found side by side with angular, cuspate, and lambdate folds.

Lambdate folds may continue at depth without change of form. Cuspate folds modified by faulting along the axial plane can persist at depth with slight change of form. Most cuspate folds in other developments show a decreasing acuteness of the fold at depth; they can overlie concentric folds or complex structures with décollements. Complex structures in the cores of angular folds, and décollements beneath, have not been seen by the writer in the field area. Mathews (1947, p. 16) presumed that complex structures might underlie angular folds, but without mention of examples or further discussion. Some non-concentric folding appears to be limited in its vertical extent. This appears to be true of some small angular folds in the coal measures of the Crassier Group, the cuspate folds on the east limb of the Crassier anticline, and the accordion folds in the Coyote Creek syncline (Fig. 3). The latter are thought to be squeezed in the closing of the syncline and partly detached from the underlying beds.

Detachment planes of great areal extent are often thought to underlie folded and thrust faulted belts. Muller (1961) applied such interpretation to the Pine River Foothills; deformed and folded Mesozoic strata of the Inner Foothills and nearby parts of the Outer Foothills are supposed to be separated from an almost undisturbed substratum of Mesozoic, Palæozoic, and basement rocks, along a single detachment plane. However, strata in the eastern part of the Inner Foothills and the adjacent border of the Outer Foothills do not compose an overthrust sheet; they are autochthonous from all the evidence at hand (*see* Appendix 2).

Little direct evidence of the relations of particular folds to underlying structures is obtained in the mapping. The Pine Valley exposes different levels of folding and close associations of concentric and angular fold forms. Associations and combinations of concentric, angular, and disharmonic folding are to be expected in the subsurface.

APPENDIX 2

THE PEACE RIVER STRUCTURES OF THE OUTER FOOTHILLS

INTRODUCTION

The structural pattern in the Outer Foothills resembles many others described in the literature. The common features are the different fold amplitudes of synclines and anticlines, the restriction of deformation to narrow anticlinal and fault zones, and the prevalence of parallel folding. Stille (1917) used the term " Injektivfaltung " to describe folding of different amplitude in synclines and anticlines in the Hannoverian Plain. This folding is considered to result from injective and diapiric movement in underlying salt beds, with or without tectonic compression (*see also* Barton, 1925, p. 439; Trusheim, 1960, p. 1519). The term " disjunctive " was given for such folding by Goguel (1952, p. 127), but without regard to the fold mechanisms. Similar patterns occur in foreland areas, the Palembang Basin (Lees, 1952, p. 14) and the foreland of the Mackenzie Mountains (Goodman, 1954, p. 352); these authors explain the deformation by folding and faulting in the basement.

Comparisons of fold patterns do not imply close similarity in all their features. The same mechanisms of folding are not indicated. For this reason the structures of the Outer Foothills are best given a local name, " the Peace River structures."

REVIEW

Several authors noted the alternation of broad synclines with narrow close-folded anticlines in the Peace River structures. McLearn (1940A, p. 74) offered a suggestion that they were due to thick competent strata, which acted as struts and so transmitted thrust forces for long distances; strata in the synclines were thought to represent the former struts. In a brief discussion, Beach and Spivak (1944, p. 11) concluded that the " folds (of the anticlines) are more superficial than folds of similar size in the southern Foothills."

Goodman (1954, p. 352) cautioned against accepting the anticlines of the Peace River Foothills as superficial folds. He suggested that they may be located approximately over basement faults, following the argument: " if these structures are interpreted as surface effects which do not continue to depth, it necessitates the assumption that the intense stress necessary to form these structures was transmitted through the weak rocks of the flat synclines without causing even a minor flexure in these beds. This appears very improbable."

The arguments concerning strength, either the individual test strengths of rocks or the mass strength of heterogeneous strata are probably unreal. All that can be said is that the rocks of the synclines and anticlines were strong enough to transmit some proportion of the stresses which deformed them. The sedimentary succession of the Foothills includes both competent and incompetent formations. The same rocks, carbonates, shales, sandstones and conglomerates, and coal measures are undisturbed or little deformed in the synclines, but bent and broken in the anticlines.

EXAMINATION OF THE STRUCTURES

The deformation of the Outer Foothills was mostly restricted to the folding and faulting of the long anticlines, as shown in Figure 25. The folding of the anticlines was concentric or angular, or both combined, for example, the Commotion, Folded

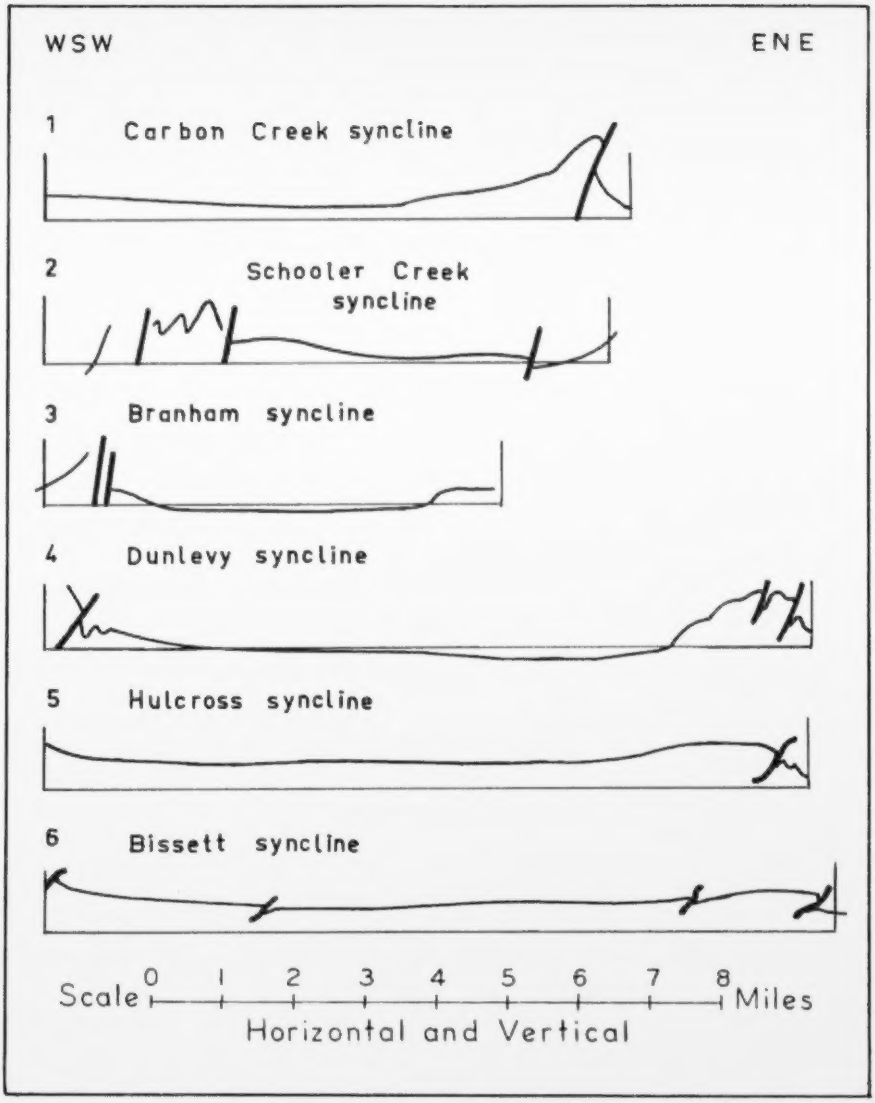

WSW
ENE
1 Carbon Creek syncline
2 Schooler Creek syncline
3 Branham syncline
4 Dunlevy syncline
5 Hulcross syncline
6 Bissett syncline
Scale 0 1 2 3 4 5 6 7 8 Miles
Horizontal and Vertical

Figure 25. The Peace River structures illustrated from cross-sections. Section 1 from Mathews (1947); 2 and 3 from McLearn (McLearn and Kindle, 1950); 4 from Beach and Spivak (1944) (*see* Table III); 5 and 6 by writer.

Hill, and Mount Gething-Stott Creek anticlines (Chapter IV; McLearn and Kindle, 1950; Beach and Spivak, 1944). Asymmetry of the folding is common, but variable from extreme to slight; the axial planes dip southwest. Much of the concentric folding can be traced to fold centres at shallow levels. The non-concentric folding may continue indefinitely in depth.

Synclines of the Peace River structures are about four times as wide as the anticlines, and have broad simple forms and slight fold amplitudes. Their west limbs are short, with dips of 25 degrees and less, decreasing eastwards. The east limbs are longer, and have dips of 5 degrees or less to an inflection of strata near the anticlines. The middle parts of the synclines are occupied by flat-lying or slightly flexured beds. The description " flat synclines " (Goodman, 1954, p. 352) is appropriate. In addition, some synclines contain subordinate folds and faults of minor scale, for example, a probable fault mapped in the Bissett syncline, folds and thrust faults about the axis of the Carbon syncline shown by Mathews (1947, p. 22).

The synclines are major structures of great length, continuous from the Peace to the Pine River. Their continuity and dimensions are listed in Table III.

Table III.—Continuity and Dimensions of Synclines in the Outer Foothills of the Peace, Moberly, and Pine River Areas

Peace River Foothills		Width (Miles)	Pine River Foothills	Width (Miles)	Length (Miles)
North	South				
Schooler, A E	Carbon, B	5 to 6	Ending against Moberly culmination[1]	—	45
Branham, A E	Branham	4 to 5	Hulcross (=Pine River), C	5 to 10	25 to 30
Dunlevy, D E	Dunlevy	7	Hulcross (=Pine River), C	5 to 10	About 75
			Bissett	8	25
Peace River Plains					
Starfish		6	Continuity uncertain	—	9

[1] A structural culmination at the Carbon Creek-Moberly River divide, in the area of the Beattie Peaks, Mounts Monteith and Frank Roy.

Data from the following authors: A, McLearn (in McLearn and Kindle, 1950, Fig. 8); B, Mathews (1947); C, Wickenden and Shaw (1943) and Spivak (1944); D, Beach and Spivak (1944); E, Irish (1961); also from observations by the writer.

In the Moberly River area, the synclines contain some oblique anticlines and flexures with faulting. These oblique structures terminate within the synclines and are subsidiary to the main longitudinal fold pattern. Two fold systems are then recognized in the Outer Foothills—the main longitudinal fold system trending south 18 degrees east to south 22 degrees east, and the subordinate, oblique Moberly fold system trending south 64 degrees east to south 44 degrees east. The Peace River structures are also found in the west margin of the Plains, adjoining the Foothills, but the deformation here was on a lesser scale.

The synclines of the Peace River structures are relatively depressed in order, from west to east across the Foothills and the nearby Plains (Fig. 3). Movement between the synclines took place across the anticlines and their thrust faults. It consisted of: (1) the relative vertical displacements of the synclines; (2) compression of the anticlines; (3) movement on the thrust faults of the anticlines. Movement between the synclines can be equated to the dip slip of a reverse fault, or thrust. The synclines can then be regarded as thrust blocks.

The movement in the anticlinal zones is referred to three structural components, namely: (1) the anticlinal thrust fault observed in mapping; (2) the fold of the anticline; (3) the underlying structure. The structure underlying the anticlines is unknown. It may take one of several forms: one fold, or a set of folds; one or more décollements providing for the anticlinal folding at the surface; a thrust of steep dip; or part of a flat thrust of extensive translation.

POSSIBLE MECHANISMS AND AVAILABLE INTERPRETATIONS

The Peace River structures may result from different geological mechanisms, and the following interpretations may be allowed for them.

The synclines form flat thrust blocks, and are separated by a structural discontinuity. The movement between the thrust blocks was reverse. It produced the folding of the anticlines and the deflections of the margins of the thrust blocks, giving them their synclinal form. Structures below the anticlines may consist of simple thrusts and folds related to thrusting. The movement between the synclines was limited, and about the same at surface as at depth; the vertical displacement between synclines is uniform at all levels. Compression and uplift were transmitted in the basement or in sediments overlying a décollement.

The Peace River structures can be interpreted as flat overthrust sheets, as for the mechanism of sheet thrusting proposed by Rich (1934, p. 1580). The anticlines are produced by bending of strata over steps of the thrust plane and footwall. The anticlinal faults connect with flat planar thrusts which cut strata in the synclines. Thrust planes in the subsurface of the synclines have not been proven or disproven. A similar condition may be produced by gravitational gliding.

Thirdly, salt tectonics may account for the Peace River structures in one of the following ways. (1) Salt movement produced the anticlines by injection. The movement resulted from isostatic adjustment and lateral compression combined. (2) A secular movement and growth of salt lenses occurred over a long period of time, and developed anticlines. This movement produced a recurrent uplift of the anticlines, accompanied by erosion of their covers. The anticlines constituted flaws along which close folding was realized in later orogenic compression. This mechanism has been considered for structures which resemble those of the Outer Foothills by Dubar (*see*, in Joly, 1952, p. 73) and Kranck (1961, p. 439). The salt provided a complete and extensive décollement, over which the sedimentary cover was moved and deformed. The movement and folding developed under compression transmitted in the sedimentary cover, as in the mechanism proposed by Buxtorf for the Jura (1916, p. 185). Otherwise the compression and movement were developed in the basement as considered by Aubert (1949, p. 14) and Glangeaud (1949, p. 686).

EVIDENCE CONCERNING THE PEACE RIVER STRUCTURES

The evidence concerning the Peace River structures comes mostly from surface mapping.

Flat sheet thrusts or repetition of strata have not been found in the synclines, and the interpretation of flat overthrusting is not indicated from mapping. If overthrusts exist, they lie concealed below river level. The Peace River structures can be seen along the main transverse valleys, in sections of 2,000 feet elevation, up to 4,000 feet in the anticlinal zones. The stratigraphic section exposed in the Foothills is about 15,000 feet thick, of which 6,000 feet is the maximum observed in one structure. Deep wells drilled on anticlines in the Outer Foothills do not suffice to prove extensive sheet thrusting (Chapters IV and V). A few anticlines end along their strike with simple plunge and decrease of their fault displacements. The Peace River structures do not include cross faults, or wrench faults of lateral movement, common adjuncts of sheet thrusting. The oblique folds along the Moberly Valley are anticlines of low amplitude and resemble the Commotion anticline.

There is no positive evidence of gravitational gliding. No important inversions of strata are present in the Outer Foothills. Chaotic or irregular structures are absent. No tears or rifts as produced in the rear of sheets of gliding and translation have been seen. The faulting along the structural trends is reverse in the anticlines mapped at the surface. The Crassier anticline, a component of the Pine River anticline along the boundary of the Inner and Outer Foothills, may allow an exception (*see* Hunt Sands Sun Falls c-18-G, Chapter V). The open Carbon syncline of the Outer Foothills passes to close folded structures of the Inner Foothills along the same structural trend at the Moberly culmination. (Table III; *see* Mathews, 1947, map.)

It can be presumed that evaporites of the Charlie Lake Formation underlie the Foothills. They consist of thin anhydritic sequences in the west part of the Pine Valley. Very little of the formation has been drilled on the east. No other evaporites are known. Salt cores have not been found in the anticlines of the Outer Foothills.

There is no evidence of major salt tectonism in the Peace River structures. Small movements in salt deposits perhaps contributed to the anticlinal folding. The concentrically folded anticlines are superficial structures—a requirement of the geometry of their folding. Hypotheses of gravitational gliding in sheets may not be discounted. Folding in the anticlinal zones and the relative vertical displacements of the synclines indicate compression and uplift at depth, with movement in the basement and overlying sediments, or in the sedimentary cover by itself.

COMPARATIVE STRUCTURES AND TECTONIC ENVIRONMENTS

Uncertainty of the role of the basement in paratectonic folding is implied in comparisons of the Peace River structures. Deformation of the sedimentary cover by movement in the basement has been advocated and disputed by many authors. Rodgers (1949) summarized the conflicting views for the Appalachians, Goodman (1962) for the Rocky Mountains (Canada), and Osterwald (1961) for the foreland of the Western Cordillera (U.S.A.). Buxtorf (1916) and Glangeaud (1949) represented different views of the Jura folding—a lateral compression of sediments above an extensive décollement and the passive role of the basement versus a deformation applied to the basement and the sedimentary cover. Usually the deformation in the basement cannot be observed directly: seismic and geophysical surveys are open to interpretation and uncertainty.

However, sufficient evidence of the deformation and tectonism of the crystalline basement exists. The subject was examined by Hudson (1955), who showed folding in the basement affecting overlying sediments; the basement rocks include granites, gneiss, and a complex of gneiss, schists, and intrusives. Structures formed by combined folding and thrust faulting of the basement and overlying sediments were reported from the Bighorn Mountains by Bucher (1933, p. 170), Wilson (1934, p. 501), and Darton (1906, p. 91). Berg (1962, p. 2022) supplied three more examples from Wyoming (*see also* Osterwald, 1961).

Comparisons of the Peace River structures can be based on the following points:—

(1) Deformation by compression.

(2) Restriction of deformation to long lines.

(3) Deformation consisting of thrust (reverse) faulting, or the folding of anticlines; these are considered tectonic equivalents.

(4) Little or non-deformed blocks, or flat synclines between the anticlines or faults.

(5) An incongruence of fold forms where folding is prevalent, and different fold amplitudes of synclines and anticlines.

A pattern of flat synclines alternating with narrow anticlines, or lines of thrust faults, is found in planar overthrust sheets as shown in the Cumberland Plateau overthrust (Rodgers, 1950; Stearns, 1954). These structures were ascribed to thrusting confined to the sedimentary cover (Rich, 1934); Harris and Zietz (1962) believed in the possibility that the deformation involved the basement. Many overthrusts share the same tectonic environment as the Peace River structures, and were formed on the cratonic edge of folded mountain belts. Extensive overthrusts may include two types. In the first, the overthrust part is undeformed or slightly warped and folded along a few lines; examples include the Cumberland overthrust; part of the McConnell overthrust as shown by Clark (1954); the Savanna Creek structure described by Scott (1953, p. 134) and in Hume (1957, p. 403). Other overthrusts are marked by continuous folding, as in the Bannock overthrust (Mansfield, 1952, p. 73). The first type is considered in comparison with the Peace River structures.

The nature of monoclinal folding in the Colorado Plateau recalls the Peace River structures. Deformation localized along fold lines and the vertical displacement of strata between the folds are common features, which permit the comparison. Several monoclines pass to thrusts, and this folding was thought to be the result of compressive stress by Kelley (1955, p. 799).

Much tectonism of the basement reproduces the structural pattern of the Outer Foothills, namely, fault blocks, of flat synclinal form, separated by long anticlinal folds or faults, mostly reverse faults. Lees (1952) collected and described examples of the concurrent deformation of the basement and its sedimentary cover, and structures wherein such deformation can be interpreted with reliance, or understood from the results of drilling. (In other examples quoted by Lees—such as the Turner Valley and Savanna Creek structures, Western Canada—he interpreted the basement structures freely.)

In structures about the Gulf of Suez, several anticlines expose a long, narrow core of the basement. The anticlines are faulted, and separate broad, shallow synclines 5 to 15 miles wide. The attitudes of the sedimentary cover and basement are observable from exposures (Lees, 1952, p. 10). The Palembang Basin, Sumatra, contains about 10,000 feet of sediments which were deformed with the basement. The deformation took the form of long anticlines with some faulting. The anticlines alternate with flat, broad synclines. This area has been explored by drilling to the basement (*see* Barnwell in Lees, 1952, p. 14).

Beckwith (1938, 1942) showed other examples from the Laramie Basin and the Laramie Valley, Colorado and Wyoming. The basement and its sedimentary cover, 10,000 to 15,000 feet thick, were folded and thrust faulted together. The long anticlines and thrusts separate fault blocks which were folded in flat, open synclines 2 to 5 miles wide. The synclines are depressed relative to each other, but a distinct unilateral step pattern is absent.

Patterns of structural steps bounded by thrusts and anticlinal folds were reported from the North Pyrenees by Destombes (1948) and from the Bergamasque Alps by de Sitter (1949). De Sitter named this pattern of structures "Le Style Nord-Pyrénéen," and indicated its development in the flanks of mountain belts and independence of gravitational gliding.

Fold and tectonic patterns comparable to the Peace River structures occur in the cratonic margins of mountain belts, and in areas of stable-metastable basement affected by orogenic stress. These environments include the forelands in the restricted and broader sense of the term.

APPENDIX 3

BRITISH COLUMBIA GOVERNMENT PINE RIVER No. 1 WELL

Location: Southwest corner of the Southwest Quarter of Section 35, Township 76, Range 26, west of the 6th Meridian. Lot 1122, north 743 feet, west 1,941 feet of the southeast corner. Approximate latitude and longitude 55° 36′ north, 121° 54′ west.

Elevation: Rotary table, 2,010 feet; ground level, 2,002 feet (Geological Survey of Canada).

DESCRIPTION OF DRILL CUTTINGS

Interval (Ft.)	Unconsolidated Rocks (Recent and Pleistocene)
20 to 40	Coarse sands and gravels: river alluvium.
40 to 70	Samples missing.
70 to 150	Clay, light grey, greenish, finely silty and micaceous, somewhat calcareous, unlaminated.
150 to 190	Clay, etc., as above; may include minor sandy pebbly beds.
190 to 850	Clay, light grey and greenish-grey, similar to above; silty and sandy to minor degree. (It is not certain whether the sand and clastic grains are in proper stratigraphic place in the clay, or whether they were embedded in the clays during drilling.)
850 to 860	Clay, light grey, greenish, calcareous, silty carrying coarse sand grains and coalified plant material.
860 to 870	Sands, coarse grained, including few clastic limestone grains.
870 to 1,081	Clay, light grey, variously greenish, blue, or brownish; somewhat calcareous, slightly silty and sandy, carrying clastic chert and quartz grains, and coalified plant remains.
	FORT ST. JOHN GROUP
	1,081—*Moosebar Formation and ?lower part of Commotion Formation.*
1,081 to 1,150	Mudstone, somewhat calcareous, with minor silty shales and siltstones.
1,150 to 1,260	Mudstone, dark grey, mostly non-calcareous; with minor siltstones; shaly.
1,260 to 1,310	Mudstone and shale, etc., as above for interval 1,150 to 1,260 ft.; with lesser shales, fissile, flaky, soft, grey, non-calcareous.
1,310 to 1,320	Shales, dark grey, and shales, grey fissile, as for intervals 1,260 to 1,310 ft. Fragments of coaly material either as cavings or drillings samples. Considerable cavings.
1,320 to 1,340	Shales, etc., as for 1,310 to 1,320 ft.; similarly, with minor sandstones and free clastic chert and quartz grains. Cavings.
1,340 to 1,500	Shales, dark grey and grey, as above; with sand or poorly consolidated sandstone, sandstones medium to very fine grained, and somewhat calcareous.
1,500 to 1,522	Mudstones and shales, dark grey, silty, mainly non-calcareous; with lesser sandstones, fine and very fine grained, partly calcareous.
1,522 to 1,530	Samples missing.
1,530 to 1,540	Mudstones, dark grey, non-calcareous.
1,540 to 1,550	Samples missing.
1,550 to 1,850	Mudstones and shales, dark grey, non-calcareous.
1,850 to 1,900	Mudstones and shales, as for interval 1,500 to 1,850 ft., above; variably and slightly calcareous in part.
1,900 to 2,010	Mudstones and shales, etc., as for the interval 1,850 to 1,900 ft.; with minor siltstones and few fine- and very fine-grained sandstones.
2,010 to 2,050	Mudstones, dark grey, variably calcareous and non-calcareous.

Interval (Ft.)	Unconsolidated Rocks (Recent and Pleistocene)
2,050 to 2,180	Mudstones, dark grey, mostly rather calcareous; with minor lighter coloured, calcareous siltstones, and very fine-grained sandstones; some glauconitic beds, notably at 2,070, 2,080, and 2,120 ft.
2,180 to 2,410	Mainly dark-grey calcareous shales and mudstones; with non-calcareous mudstones and shales of similar appearance; some glauconitic shales, from 2,400 ft. to base of formation.
2,410 to 2,424	Mudstones and shales as for interval 2,180 to 2,410 ft. Free clastic chert grains and fragments, coarse and very coarse. Carbonaceous shale, coals and dark grey siltstones.

CRASSIER GROUP

2,420—*Gething Formation and ?lower parts of Crassier Group.*

2,424 to 3,330	Shales and mudstones, dark grey; carbonaceous shales and silty mudstones; siltstones; few sandstones, mostly very fine grained; coals.
3,330 to 3,640	Shales and mudstones, dark grey; carbonaceous shales and mudstones; silty mudstones; siltstones; sandstones, mostly fine to very fine grained; coals.
3,640 to 3,970	Shales and mudstones, dark grey; carbonaceous shales and mudstones; silty mudstones; siltstones, sandstones, fine and very fine grained; with much medium-grained sandstones, grey; coals.
3,970 to 3,980	Sandstones, coarse grained; few cuttings of loose chert granules.
3,980 to 4,740	Shales and mudstones, dark grey; carbonaceous shales and mudstones; silty mudstones; sandstones, medium grained, grey, or speckled with dark-coloured chert grains; coals.
4,740 to 5,040	Sandstones, mostly very fine and fine grained, and lesser medium-grained sandstones, grey. Lesser shales, mudstones and coals.
5,040 to 5,390	Shales and mudstones, dark grey; silty mudstones; and carbonaceous mudstones; siltstones; sandstones, very fine and fine grained; some medium-grained sandstones, grey or speckled; coals.
5,390 to 5,410	Sandstones, fine and medium grained, grey and speckled; some shales, mudstones, and siltstones.
5,410 to 5,420	Sandstones, fine to very coarse grained, and grits, speckled; cuttings of chert and quartzite granules, and pebbles.
5,420 to 5,450	Siltstones, grey; shales and mudstones; carbonaceous shales and coals; lesser sandstones, mostly fine grained.
5,450 to 5,480	Sandstones, fine to very coarse grained, speckled; cuttings of chert and quartzite pebbles and granules in upper part. Some shales, mudstones, siltstones, and coals.
5,480 to 5,540	Shales and mudstones, dark grey; carbonaceous shales and mudstones; silty mudstones; minor sandstones, fine and medium grained; coals.
5,540 to 5,590	Sandstones, fine to very coarse grained and grits, speckled; cuttings of chert and quartzite pebbles and granules.
5,590 to 5,730	Shales and mudstones, dark grey; carbonaceous shales and mudstones; silty mudstones; siltstones, sandstones, medium to coarse grained, grey or speckled; few grits and few cuttings of chert and quartzite granules; coals.
5,730 to 5,960	Shales and mudstones, dark grey; carbonaceous and silty mudstones; abundant siltstones; sandstones in lesser amounts and mostly fine-grained sandstones, with few medium-grained, grey sandstones; coals.
5,960 to 6,040	Coals mostly.
6,040 to 6,260	Shales and mudstones, dark grey; carbonaceous shales and mudstones; silty mudstones; siltstones; sandstones, very fine to fine grained, grey; coals.
6,260 to 6,290	Sandstones, medium grained, speckled; few coarse-grained sandstones and cuttings of chert granules; minor sandstones, fine grained, and shales.
6,290 to 6,400	Shales and mudstones, dark grey; carbonaceous shales and mudstones; silty mudstones; siltstones; sandstones, very fine and fine grained; coals.
6,400 to 6,941	Shales and mudstones, dark grey; carbonaceous shales and mudstones; silty mudstones; siltstones; sandstones, very fine and fine grained, medium grained, grey; coals.

(*Total depth:* 6,941 ft., driller.)

DESCRIPTION OF CORES

CRASSIER GROUP

Interval (Ft.)	Recovery		Description
	Ft.	In.	
Core No. 1—2,915 to 2,918½	3	4	Shale, dark grey, silty; with thin interbedded and interlaminated siltstones.
Core No. 2—3,146 to 3,149	0	4	Mudstone, dark grey, compact.
	0	9	Siltstone, dark grey, shaly; showing finely marked small scale current bedding.
	0	8	Shale, with thin coaly partings.
Core No. 3—3,908 to 3,909	0	4	Sandstone, grey, medium grained, clastic chert grains.
	0	4½	Shale, dark grey, carbonaceous with few plant fragments.
	0	2½	Shale, grey, silty; with interlaminated siltstones.
Core No. 4—4,662 to 4,664	2	0	Siltstones, dark grey; with finely interlaminated dark grey shales.
Core No. 5—4,686 to 4,888½	1	6	Sandstone, coarse and medium grained, pebbly, gritty feel; speckled appearance, with dark-coloured clastic chert grains.
Core No. 6—4,740½ to 4,741½	0	5	Sandstone, grey, speckled, medium grained, chert bearing; containing fragments of shale.
Core No. 7—4,786 to 4,789	2	6	Sandstone, grey speckled, medium to fine grained, coarse bedded; chert and quartz grains.
Core No. 8—4,822 to 4,824½	1	0	Sandstone, silty and shaly, very fine grained; interlaminated with dark grey shales. Fine small scale current bedding.
Core No. 9—4,863 to 4,867	1	5	Sandstone, grey, speckled, cherty, medium to fine grained. The sandstone contains a few shaly partings and shale fragments.
Core No. 10—4,994 to 4,995½	0	9	Sandstone, grey, speckled, cherty.
Core No. 11—5,178 to 5,182	0	10	Mudstone, dark grey, carbonaceous.
	3	9	Mudstone, dark grey, carbonaceous, etc., as above; considerable disperse, fine-veined calcite.
Core No. 12—5,353 to 5,358½	2	9	Mudstone, dark grey to black.
Core No. 13—5,522 to 5,525	2	9	Mudstone, dark grey, carbonaceous, with coalified plant fragments.
Core No. 14—5,732 to 5,736	4	2	Mudstone, dark grey, carbonaceous, similar to Core No. 13 above.
Core No. 15—5,901 to 5,906½	5	10	Sandstone, very fine grained; siltstones, dark grey; with interlaminated shales; some small scale current bedding.
Core No. 16 Core No. 17 Core No. 18	—	—	Cores missing.
Core No. 19—6,249 to 6,252	2	6	Siltstone, grey, shaly, tough; showing fine lamination and current bedding.
Core No. 20—6,324 to 6,327	2	0	Sandstone, grey, fine grained, silty, finely cross-bedded; with interlaminated shales.
Core No. 21—6,488 to 6,490	2	0	As above for Core No. 20.
Core No. 22—cut about 6,700 ft.	2	4	Shale, black, carbonaceous; with fragments of coalified plant material.
Core No. 23—6,848 to 6,853	2	6	Mudstone, black, carbonaceous, silty, in places, and with fine silty partings; few thin coaly partings, and some traces of plant fragments; containing single sandstone parting ¼ in. thick.

RECORD OF DIPS OBSERVED IN CORES

Interval (Ft.)	Dip (Deg.)
Core No. 1—2,915 to 2,918½	26 to 35
Core No. 2—3,146 to 3,149	5
Core No. 3—3,908 to 3,909	3 to 8
Core No. 4—4,662 to 4,664	75
Core No. 5—4,686 to 4,688½	90 vertical
Core No. 6—4,740½ to 4,741½	65
Core No. 7—4,786 to 4,789	75
Core No. 8—4,822 to 4,824½	70 to 90 to 110. The core shows a small flexure or dragfold.
Core No. 9—4,863 to 4,867	85 to 90 vertical
Core No. 10—4,994 to 4,995½	70 to 73
Core No. 11—5,178 to 5,182	Mudstones with no clear bedding; dips are indeterminate.
Core No. 12—5,353 to 5,338½	
Core No. 13—5,522 to 5,535	
Core No. 14—5,732 to 5,736	
Core No. 15—5,901 to 5,906½	83
Core No. 16	Cores missing.
Core No. 17	
Core No. 18	
Core No. 19—6,249 to 6,252	5 to 10
Core No. 20—6,324 to 6,327	20 to 25
Core No. 21—6,488 to 6,490	30
Core No. 22—Core cut at about 6,700 ft.	5 to 10
Core No. 23—6,848 to 6,853	15 to 25

RECORD OF DEVIATION OF HOLE

Depth (Ft.)	Deviation from Vertical (Deg.)	Depth (Ft.)	Deviation from Vertical (Deg.)
1,200	⅙	5,580	7¾
2,080	1⅓	5,680	9½
3,080	1¾	6,000	8½
4,000	2	6,200	7
4,800	3½	6,400	7¼
5,410	3	6,600	9
5,520	4¾	6,800	10

The record of deviations is taken from a compilation of data concerning the well by T. B. Williams (1942).

APPENDIX 4

SUN ET AL CHETWYND 14-20-77-23 WELL

Location: 578 feet south and 658 feet west from the northeast corner of Legal Subdivision 14, Section 20, Township 77, Range 23, west of the 6th Meridian.
Elevation: Rotary table, 2,669 feet; ground level, 2,653 feet.

DESCRIPTION OF DRILL CUTTINGS

DUNVEGAN FORMATION +1,194 Ft.

Interval (Ft.)	
0 to 10	Sandstones, very fine grained.
10 to 20	Shales, light coloured and grey.
20 to 30	Shales, light coloured and grey; carbonaceous shales. Siltstones. Coals, minor.
30 to 50	Sandstones, very fine to fine grained. Shales and silty shales, light coloured, grey, red, and brown.
50 to 520	Shales, light coloured, red, brown; shales and silty shales, dark grey; carbonaceous shales, and shales with plant and coaly debris. Siltstones. Lesser sandstones, very fine to medium grained, with chert, and lithic grains. Coals in minor proportions at 320, 340, and 450 ft.
520 to 550	Sandstones, very fine to fine grained, micaceous; and with chert and lithic grains, and coaly fragments—as for Dunvegan sandstones generally. Siltstones. Coals, minor.
550 to 750	Shales, light coloured, grey, and dark grey, silty shales, dark grey; carbonaceous shales, and shales with plant and coaly debris. Siltstones. Lesser sandstones, very fine grained. Flakes of muscovite, common at 630, 640, and 670 feet. Coals in minor proportions at 630 feet, and from 560 to 650 feet (?part cavings).
750 to 770	Sandstones, very fine to fine grained. Lesser shales, as for 550 to 750 feet. Cuttings of chert. Flakes of muscovite. Coals, ?cavings.
770 to 780	Shales, with lesser siltstones and sandstones, as for 550 to 770 feet.
780 to 830	Sandstones, very fine to medium grained, as for 520 to 550 feet. Lesser shales, as for 550 to 770 feet. Siltstones.
830 to 940	Shales, as for 550 to 770 feet. Lesser siltstones, and sandstones. Coals, minor—cavings or cuttings undistinguished.
940 to 980	Mostly sandstones, very fine to fine grained. Lesser shales, as for 550 to 770 feet. Cuttings of chert, loose chert and quartz grains in lower part. Flakes of muscovite. Coals, minor—cavings or cuttings undistinguished.
980 to 1,010	Shales, as for 550 to 770 feet. Few siltstones, and sandstones.
1,010 to 1,020	Sandstones, very fine to fine grained. Cuttings of chert. Few shales.
1,020 to 1,090	Shales, as for 550 to 770 feet. Lesser siltstones, and sandstones. Cuttings of chert, at 1,040 feet. Coals at 1,050 feet, minor—cavings or cuttings undistinguished.
1,090 to 1,110	Sandstones, very fine to fine grained; and micaceous sandstones.
1,110 to 1,120	Siltstones, grey, and dark grey. Shales, and silty shales, dark grey. Lesser sandstones, very fine grained.

1,120 Lower (or *Unio* (*P.*) cf. *dowlingi*) Beds 90 Ft.

1,120 to 1,210	Siltstones, grey, and dark grey; micaceous and argillaceous siltstones; siltstones in decreasing proportions downward. Shales, and silty shales, dark grey. Lesser to few sandstones, very fine grained, micaceous, and argillaceous. Few shell fragments at 1,150 feet.

FORT ST. JOHN GROUP

1,210 *Cruiser Formation* 632 Ft.

1,210 to 1,230	Shales, with silty shales, dark grey.
1,230 to 1,360	Shales, dark grey; few silty shales. Siltstones—minor at 1,330 feet.
1,360 to 1,660	Shales, with silty shales, dark grey. Few siltstones. Minor sandstones, very fine grained. (Samples missing at 1,420 and 1,430 ft.)

Interval (Ft.)	
1,660 to 1,690	Sandstones very fine grained, clean, with siliceous matrix. Shales, and silty shales, dark grey.
1,690 to 1,710	Shales, with silty shales, dark grey.
1,710 to 1,730	Shales, dark grey. Minor siltstones, and sandstones, very fine grained.
1,730 to 1,780	Shales, dark grey—mostly.
1,780 to 1,810	Shales, dark grey. Few siltstones, and sandstones, very fine grained.
1,810 to 1,830	Shales, dark grey.

1,842 *Shales with Sandstones* 760 Ft.

1,842 Fourth Arenaceous Member 38 Ft.

1,830 to 1,860	Sandstones, mostly very fine to fine grained, clean; containing chert grains, lesser lithic grains. Minor glauconitic sandstones. Clean sandstones, and argillaceous sandstones, and shales in lower 10 feet.
1,860 to 1,870	Shales, dark grey; few silty shales. Few sandstones.
1,870 to 1,880	Sandstones, very fine to fine grained; pyritic sandstones. Shales, dark grey. Siltstones.

1,880 Third Argillaceous Member 279 Ft.

1,880 to 1,910	Shales, dark grey; few silty shales. Few sandstones, argillaceous sandstones, and siltstones.
1,910 to 1,920	Shales, dark grey.
1,920 to 1,970	Shales, and silty shales, dark grey. Lesser siltstones, and sandstones, fine grained.
1,970 to 1,990	Shales, dark grey; few silty shales. Minor siltstones.
1,990 to 2,040	Shales, dark grey. Minor sandstones at 2,030 and 2,040 ft.
2,040 to 2,110	Shales, dark grey; silty shales, siltstones, decreasing in lower part. Few sandstones, very fine grained, in upper part, and at 2,110 ft.
2,110 to 2,150	Shales, dark grey, for most part. (Sample missing at 2,120 ft.)

2,159 Third Arenaceous Member 59 Ft.

2,150 to 2,200	Shales, dark grey; silty shales. Lesser siltstones, and sandstones, very fine grained. Few coarse-grained sandstones and cuttings of chert granules at 2,200 ft.
2,200 to 2,220	Sandstones, very fine to medium grained, clean, with dark-coloured chert grains. Siltstones. Few cuttings of chert. Sandstones, with siltstones, and shales in lower part.

2,218 Second Argillaceous Member 97 Ft.

2,220 to 2,240	Siltstones, and argillaceous siltstones. Silty shales, dark grey. Few sandstones, very fine grained.
2,240 to 2,270	Shales, and silty shales, dark grey. Lesser siltstones, and argillaceous siltstones.
2,270 to 2,320	Shales, dark grey; few silty shales.

2,315 Second Arenaceous Member 43 Ft.

2,320 to 2,340	Sandstones, very fine to medium grained, clean, with dark-coloured chert grains. Glauconitic sandstones, as above, common. Siltstones. Silty shales, dark grey; few glauconitic shales.
2,340 to 2,360	Shales, and silty shales, dark grey. Siltstones. Lesser sandstones, very fine grained.

2,358 First Argillaceous Member 216 Ft.

2,360 to 2,460	Shales, dark grey; silty shales. Few siltstones. Minor sandstones, very fine grained. Shell fragments at 2,420 ft.
2,460 to 2,480	Shales, dark grey. Shell fragments.
2,480 to 2,570	Shales, dark grey; silty shales. Lesser siltstones. Sandstones, very fine grained, in minor and variable proportions.

2,574 First Arenaceous Member 28 Ft.

2,570 to 2,610	Shales and silty shales, dark grey. Sandstones, very fine to fine grained, clean, with chert grains. Argillaceous sandstones, and siltstones. Cuttings of chert, and chert granules at 2,580 ft.

2,602 *Hasler Formation* 785 Ft.

Interval (Ft.)	
2,610 to 2,620	Shales, dark grey, silty, and sandy. Lesser argillaceous siltstones and sandstones.
2,620 to 2,630	Shales, dark grey, silty and sandy.
2,630 to 2,880	Shales, dark grey; few silty shales, and shales with thin silty-sandy laminæ. Few siltstones. Sandstones, very fine grained in minor proportions at 2,690, 2,720, 2,750, and 2,760 ft.
2,880 to 2,920	Shales, dark grey; few silty shales, and shales with thin silty-sandy laminæ. Sandstones, very fine grained, in minor proportions.
2,920 to 3,110	Shales, dark grey, mostly. Few, silty shales, and shales with thin silty-sandy laminæ, siltstones, and argillaceous siltstones at 2,970, 3,060, 3,070, 3,100, and 3,110 ft. Sandstones, very fine grained, in minor proportions at 2,970, 3,060, and 3,070 ft.
3,110 to 3,140	Argillaceous siltstones, and siltstones. Shales, and silty shales, dark grey. Sandstones, very fine grained, in less proportion at 3,120, and in minor proportions at 3,130 and 3,140 ft.
3,140 to 3,240	Shales, dark grey, mostly. Few silty shales, siltstones, and argillaceous siltstones. Sandstones, very fine grained, in minor proportions at 3,210 and 3,220 ft.
3,240 to 3,380	Shales, dark grey, mostly. Silty shales and siltstones, minor.

3,387 *Commotion Formation* 1,417 Ft.

3,387 Members (iv and iii) 319 Ft.

3,380 to 3,390	Shales, silty shales, and siltstones, as for 3,240 to 3,380 ft. Sandstones, fine to coarse grained, clean, speckled with light- and dark-coloured chert grains. Cuttings of chert and chert granules.
3,390 to 3,420	Sandstones, fine to coarse grained, with cuttings of chert, etc., as for 3,380 to 3,390 ft.
3,420 to 3,430	Sandstones, very fine to medium grained, resembling those of 3,380 to 3,420 ft. Cuttings of chert and chert granules.
3,430 to 3,450	Sandstones, fine to coarse grained, with cuttings of chert, etc., as for 3,380 to 3,420 ft.
3,450 to 3,480	Sandstones, very fine to coarse grained, with cuttings of chert, etc., as for 3,380 to 3,450 ft. Lesser shales, silty shales, carbonaceous shales; few shales with plant and coaly debris. Siltstones. Coals, minor: ?cavings (less than 1 per cent of sample).
3,480 to 3,490	Shales and silty shales, dark grey; few carbonaceous shales and shales with coaly debris. Lesser sandstones. Few siltstones.
3,490 to 3,510	Sandstones, very fine to coarse grained, clean, light coloured. Cuttings of chert and chert grains.
3,510 to 3,520	Sandstones, as for 3,490 to 3,510 ft. Shales, dark grey.
3,520 to 3,530	Sandstones, very fine to coarse grained; with light- and dark-coloured chert grains, and lithic grains. Cuttings of chert and chert granules, and of quartz.
3,530 to 3,570	Sandstones, as for 3,520 to 3,530 ft., etc. Lesser shales, silty shales, carbonaceous shales, and shales with coaly debris. Coals, minor: ?cavings (about 1 per cent of sample at 3,540 feet, and less thereafter).
3,570 to 3,610	Sandstones, very fine to coarse grained, and cuttings, as for 3,520 to 3,570 ft. Few shales, silty shales, and carbonaceous shales.
3,610 to 3,650	Sandstone, very fine to medium grained, light coloured and speckled; with chert and lithic grains. Ironstone, minor, at 3,640 and 3,650 ft.
3,650 to 3,720	Sandstones, very fine to fine grained, for most part, and resembling those of 3,610 to 3,650 ft. Lesser sandstones, very fine grained, grey; and sandstones with shale laminæ. Few coarser-grained sandstones, with cuttings of chert and chert granules, at 3,700 and 3,710 ft. Siltstones. Shales, and silty shales, dark grey. Ironstone.

3,706 Member (ii) 404 Ft.

3,720 to 3,800	Shales and silty shales, dark grey. Siltstones and argillaceous siltstones. Minor sandstones, very fine grained, at 3,790 ft. Mnior pyrite, in places. Glauconite in few siltstones and argillaceous siltstones at 3,780 ft.

Interval (Ft.)	
3,800 to 3,830	Shales, dark grey; minor silty shales.
3,830 to 3,850	Shales and silty shales, dark grey. Siltstones.
3,850 to 3,980	Shales, dark grey; minor silty shales. Siltstones and much silty shales at 3,930 and 3,940 ft.
3,980 to 4,000	Shales, dark grey; minor silty shales. Minor siltstones. Minor sandstones, very fine and fine grained, with coarse-grained sandstones in lower part.
4,000 to 4,040	Shales, dark grey; minor silty shales. Minor sandstones at 4,020 feet.
4,040 to 4,110	Shales, dark grey; few silty shales. Few or minor siltstones.
	4,110 Member (i) 694 Ft.
4,110 to 4,120	Sandstones, very fine to fine grained, with chert and lithic grains; some glauconite, and pyrite in places.
4,120 to 4,130	Sandstones, very fine and fine grained, with chert and lithic grains. Siltstones. Shales and silty shales, dark grey.
4,130 to 4,150	Sandstones, very fine and fine grained, as for 4,120 to 4,130 feet; few coarser-grained sandstones. Cuttings of chert, and chert granules. Shales and silty shales, dark grey.
4,150 to 4,160	Sandstones, cuttings, and shales, as for 4,130 to 4,150 feet. Shales and silty shales, dark grey; few carbonaceous shales. Coals (about 5 per cent of sample).
4,160 to 4,210	Sandstones, fine and medium grained. Cuttings of chert and chert granules at 4,160 feet.
4,210 to 4,310	Samples missing.
4,310 to 4,320	Shales and siltstones, ?cavings.
4,320 to 4,340	Samples missing.
4,340 to 4,370	Sandstones, fine and medium grained. Cuttings of chert and chert granules at 4,350 feet.
4,370 to 4,410	Samples missing.
4,410 to 4,420	Shales and silty shales, dark grey. Siltstones. Minor sandstones, light coloured.
4,420 to 4,430	Sample missing.
4,430 to 4,490	Siltstone, grey. Lesser shales, dark grey. Few sandstones, very fine grained.
4,490 to 4,500	Shales, dark grey. Lesser siltstones and silty shales.
4,500 to 4,570	Shales and silty shales, dark grey. Siltstones, grey. Lesser to minor sandstones, very fine grained.
4,570 to 4,590	Sandstones, very fine grained. Shales, dark grey. Siltstones and argillaceous siltstones.
4,590 to 4,610	Siltstones, grey; argillaceous siltstones. Shales, dark grey. Lesser sandstones, very fine grained.
4,610 to 4,630	Shales and silty shales, dark grey. Lesser siltstones and argillaceous siltstones. Few sandstones, very fine grained. Glauconite, uncommon, in shales, siltstones, and sandstones.
4,630 to 4,650	Siltstones; argillaceous siltstones. Shales, silty and sandy. Sandstones, very fine grained; argillaceous sandstones. Glauconite, common, in shales, siltstones, and sandstones.
4,650 to 4,700	Sandstones, very fine to fine grained, light coloured. Sandstones, very fine grained, argillaceous. Shales and silty-sandy shales, dark grey. Glauconite: common, in argillaceous sandstones, in siltstones, and shales at 4,660 ft.; thereafter, less common, and in decreasing proportions, in sandstones, siltstones, and shales.
4,700 to 4,740	Shales and silty-sandy shales, dark grey. Lesser sandstones, fine grained. Few siltstones.
4,740 to 4,770	Sandstones, very fine to fine grained. Siltstones. Shales and silty-sandy shales, dark grey.
4,770 to 4,810	Shales and silty-sandy shales, dark grey. Lesser sandstones and siltstones.
	4,804 *Moosebar Formation* 1,083 Ft.
4,810 to 4,820	Shales, dark grey, mostly; lesser silty shales. Few sandstones and siltstones.
4,820 to 4,870	Samples missing.
4,870 to 5,110	Shales and silty shales, dark grey. Lesser or minor siltstones. Few sandstones, in places.
5,110 to 5,140	Shales, dark grey; few silty shales.
5,140 to 5,630	Shales, dark grey, mostly. Samples missing at 5,190, 5,220, 5,230, and 5,240 ft.

Interval (Ft.)	
5,630 to 5,650	Shales, dark grey. Few siltstones, and sandstones, very fine grained.
5,650 to 5,680	Shales and calcareous shales, dark grey. Minor ironstone. Few glauconitic shales at 5,670 and 5,680 ft.
5,680 to 5,700	Shales and calcareous shales, dark grey. Lesser siltstones. Few sandstones, very fine grained.
5,700 to 5,800	Shales and calcareous shales, dark grey. Lesser or minor siltstones, and sandstones, very fine grained, in places. Glauconitic shales, and arenaceous shales, at 5,760 ft.
5,800 to 5,860	Shales and calcareous shales, dark grey. Glauconitic shales at 5,820 and 5,840 ft.
5,860 to 5,880	Shales and few calcareous shales, dark grey. Siltstones and sandstones, very fine grained; in part, argillaceous, calcareous, or pyritic. Glauconitic shales and siltstones.

CRASSIER GROUP

5,887 *Gething Formation* 513 Ft.

5,880 to 5,890	Shales, dark grey; few calcareous shales. Shales with plant and coaly debris. Sandstones, very fine to fine grained. Siltstones, minor. Ironstone. Coals.
5,890 to 5,960	Shales, dark grey; carbonaceous shales; and shales with plant and coaly debris. Sandstones, very fine grained; with chert and lithic grains (mostly shale)—as for the Crassier Group generally. Siltstones. Coals, abundant: more than 20 per cent of sample at 5,940 ft.
5,960 to 5,990	Samples missing.
5,990 to 6,300	Shales, dark grey; carbonaceous shales; and shales with plant and coaly debris. Sandstones, mostly very fine grained, argillaceous or clean: sandstones predominant at 6,060, 6,090, 6,120, 6,150, and 6,160 ft. Much siltstones and argillaceous siltstones. Coals, abundant: 20 per cent or more of sample at 6,000, 6,100, 6,170, 6,210, 6,250, 6,260, 6,280, and 6,290 ft.
6,300 to 6,400	Coal measures, as for 5,990 to 6,300 ft. Sandstones, very fine grained: predominant at 6,310 and 6,330 ft. Siltstones, shales, etc. Coals, in less proportions: about 20 per cent of sample at 6,330 ft.

6,400 *Dresser Formation* 670 Ft.

6,400 to 6,440	Sandstones, very fine to medium grained, speckled, with light and dark chert grains, and dark lithic grains (mostly of shale)—as for the Dresser Formation, and Crassier Group generally. Few shales. Coals: seams indicated from cuttings at 6,430 and 6,440 ft. (coal about 15 per cent of sample).
6,440 to 6,530	Shales, dark grey; silty shales, carbonaceous shales, and shales with plant and coaly debris. Siltstones. Sandstones, mostly very fine grained. Coals.
6,530 to 6,560	Sandstones, very fine grained; with fine-grained sandstones in lower part. Few shales. Few coals.
6,560 to 6,590	Shales, dark grey; silty shales; carbonaceous shales, and shales with plant and coaly debris. Few sandstones and siltstones.
6,590 to 6,610	Sandstones, fine to medium grained; speckled; few coarse-grained parts. Cuttings and grains of chert.
6,610 to 6,680	Sandstones, very fine to fine grained; some medium-grained parts. Shales, dark grey; silty shales; carbonaceous shales, and shales with plant and coaly debris. Lesser siltstones. Coals: seams indicated from cuttings at 6,630, 6,640, and 6,670 ft. (coal about 15 per cent of sample).
6,680 to 6,710	Sandstones, very fine to medium grained; cuttings of chert. Coals: about 50 per cent of samples. Sample missing at 6,700 ft.
6,710 to 6,770	Sandstones, very fine to medium grained, speckled; lesser coarse-grained sandstones at 6,750, 6,760, and 6,770 ft. Cuttings of chert (granules and pebbles, as for overlying beds), and chert grains, more notably in lower 20 ft. Shales and siltstones of coal-measure types, increasing in proportion to lower parts. Coals: seams indicated from cuttings at 6,760 and 6,770 ft. (coal about 15 per cent of sample).
6,770 to 6,810	Shales, dark grey, carbonaceous shales and silty shales. Lesser sandstones, very fine to medium grained; few coarse-grained parts. Siltstones. Cuttings and grains of chert, at 6,790 and 6,810 ft. Coals: including 10 per cent of sample at 6,790 ft.
6,810 to 6,850	Sandstones, medium grained; fine- and coarse-grained parts. Few shales and siltstones. Cuttings and grains of chert.

Interval (Ft.)	
6,850 to 7,080	Sandstones, largely, very fine to fine grained: medium-grained sandstones, abundant at 6,930, 6,970, and 7,040 ft., and in small amounts elsewhere. Shales, dark grey; silty shales; carbonaceous shales, and shales with plant and coaly debris. Siltstones. Cuttings and grains of chert, at 6,890 and 7,000 ft. Coals: seams indicated from cuttings at 6,920, 6,980, 6,990, and 7,010 ft. (coals about 15 per cent or more of sample).
	7,070 *Brenot Formation* 305 Ft.
7,080 to 7,090	Sample missing.
7,090 to 7,160	Sandstones, shales, few siltstones, coals. Sandstones, largely very fine grained; few fine-grained sandstones at 7,120, 7,140, and 7,150 ft.; clastic chert, grains and cuttings at 7,150 ft. Shales, dark grey; carbonaceous shales; shales with plant and coaly debris; silty shales. Coals, minor: seams indicated from cuttings at 7,100 and 7,140 ft. (coals about 10 to 20 per cent of sample).
7,160 to 7,250	Sandstones, shales, much siltstones, coals. Sandstones, mostly very fine grained; few fine- to medium-grained sandstones at 7,210, 7,220, and 7,230 ft.; clastic chert, grains, and cuttings at 7,180 ft. Siltstones, clean to argillaceous. Shales, as for 7,090 to 7,160 ft. Coals, minor: seams indicated from cuttings at 7,220, 7,230, and 7,250 ft. (coals about 10 per cent or up to 20 per cent of sample).
7,250 to 7,370	Sandstones, shales, siltstones, coals. Sandstones, few: mostly very fine grained, and grading to siltstones. Shales predominant; as for 7,090 to 7,160 ft. Coals, minor: seams indicated from cuttings at 7,270, 7,310, 7,320, 7,360, and 7,370 ft. (coals about 10 per cent of sample). Sample missing at 7,330 ft.
	7,375 *Chetwynd Beds* 245 Ft.
7,370 to 7,380	Sandstones, siltstones, shales, and few coals—as for overlying interval. Quartzites and quartzitic sandstones; grains, very fine, well sorted, well rounded or with small secondary overgrowths. Lesser sandstones, very fine to medium grained, argillaceous, and speckled; with lithic grains, mostly of dark shale, some of chert.
7,380 to 7,390	Sample missing.
7,390 to 7,400	As for 7,370 to 7,380 ft.—with more sandstones, lesser quartzites and quartzitic sandstones.
7,400 to 7,450	Sandstones, grey, very fine grained and grading to arenaceous siltstones. Sandstones, very fine to medium grained, argillaceous, and speckled with lithic grains, mostly of shale, and few of chert. Shales, dark grey-black; arenaceous shale; some carbonaceous shales. Coals, ?cavings (less than 5 per cent of sample).
7,450 to 7,460	Quartzites and quartzitic sandstones, fine to very coarse grained; with much secondary quartz. Lesser sandstones, fine to coarse grained; argillaceous; with interstitial shaly material; speckled, with lithic grains of shale, fewer of chert.
7,460 to 7,510	Sandstones, very fine to medium grained, argillaceous, and speckled with lithic grains (as for 7,450 to 7,460 ft.). Lesser shales, dark grey and black. Few siltstones. Quartzites, fine to coarse grained, in minor amounts. Cuttings and grains of quartz. Coals, ?cavings (about 3 per cent, or less of sample).
7,510 to 7,550	Samples missing—cored interval.
7,550 to 7,620	Sandstones, very fine to medium grained, argillaceous, and speckled with lithic grains (as for 7,450 to 7,460 ft.). Lesser shales, dark grey and black; carbonaceous shales, and shales with plant debris. Few siltstones. Coals: in notable amounts, at 7,560 ft. (10 to 20 per cent of sample), and 7,570 ft. (8 to 12 per cent of sample); in ?cavings, from 7,570 to 7,620 ft. (about 5 per cent or less of sample). Sample missing at 7,610 ft.
	BEAUDETTE GROUP
	7,620 *Beaudette Group Undivided* 688 Ft.
7,620 to 7,640	Shales, dark grey to black, uniform. Lesser sandstones: sandstones, very fine grained; few sandstones, very fine to medium grained, argillaceous and speckled (as for 7,450 to 7,460 ft.). Coals, ?cavings (insignificant proportions of samples).

Interval (Ft.)	
7,640 to 7,700	Sandstones, mostly very fine to fine grained, argillaceous, and speckled. Shales, dark grey to black; silty shales. Lesser or minor quartzites and quartzitic sandstones, fine to coarse grained, with much secondary quartz. Cuttings and grains of quartz, few of chert. Coals, probably cavings (insignificant proportions of samples).
7,700 to 7,730	Sandstones, very fine to coarse grained, with much quartz in secondary overgrowths, argillaceous with much interstitial shaly material, and speckled with lithic grains, mostly of dark shale. Few quartzites. Cuttings and grains of quartz.
7,730 to 7,870	Shales, dark grey to black; silty shales; and few shales with carbonaceous laminæ. Sandstones, very fine to medium grained, argillaceous, as for 7,700 to 7,730 ft. and overlying intervals; in places, gradational to siltstones and arenaceous shales. Siltstones and argillaceous siltstones. Few quartzites, and quartzitic sandstones. Cuttings and grains of quartz. (Sandstones common at 7,730, 7,780, 7,820, 7,860, and 7,870 ft.) Sample missing at 7,830 ft.
7,870 to 7,970	Sandstones, very fine to fine grained, argillaceous; with shale matrix, and lithic grains of dark shale; lesser sandstones, medium grained, as for overlying intervals. Shales, dark grey and black; arenaceous shales. Siltstones and argillaceous siltstones. Few quartzitic sandstones. Cuttings and grains of quartz at 7,880 and 7,890 ft.
7,970 to 7,980	Sample missing.
7,980 to 8,010	Sandstones, very fine to fine grained, clean, and variably argillaceous and speckled. Shales, dark grey and black; silty shales.
8,010 to 8,040	Shales, dark grey and black; silty shales. Siltstones. Lesser sandstones, clean, and variably argillaceous and speckled. Few quartzitic sandstones, very fine to fine grained. Few cuttings and grains of quartz, and lesser of chert.
8,040 to 8,210	Sandstones, very fine to medium grained, clean, and variably argillaceous; much secondary quartz; argillaceous sandstones with shaly matrices, and lithic grains mostly of dark shale, as for overlying beds. Siltstones and argillaceous siltstones. Shales, dark grey and black; silty shales. Few quartzites and quartzitic sandstones. Cuttings and grains of quartz in places. Samples missing at 8,090, 8,100, 8,180, and 8,190 ft.
8,210 to 8,290	Shales, dark grey and black; arenaceous shales. Siltstones and argillaceous siltstones. Lesser sandstones, very fine to fine grained, clean, or mostly argillaceous.
8,290 to 8,310	Shales, siltstones, and sandstones, as for 8,210 to 8,290 ft.: sandstones to about 50 per cent of samples. Sandstones, very fine to medium grained, argillaceous, with shale matrix and grains, as for overlying beds generally. Few quartzites and quartzitic sandstones, very fine to medium grained. Few cuttings and grains of quartz, less of chert.

FERNIE GROUP

8,308 *Transition Beds* 97 Ft.

8,310 to 8,410	Shales, dark grey, silty shales. Siltstones and argillaceous siltstones. Sandstones, mostly very fine grained and argillaceous (from 20 to 40 per cent, decreasing downwards).

8,405 *Middle Shales* 313 Ft.

8,410 to 8,500	Shales, dark grey and black; few arenaceous shales. Siltstones and sandstones in minor amounts. Cavings.
8,500 to 8,710	Shales, dark grey and black, mostly. Few siltstones and sandstones, undistinguished from cavings.
8,710 to 8,730	Shales, black; pyritic shales; silty and sandy shales; few calcareous shales in lower part.
8,730 to 8,760	Shales, black; pyritic shales; silty and sandy shales. Siltstones, argillaceous. Sandstones, very fine grained.
8,760 to 8,770	Shales, black; fissile and pyritic shales.
8,770 to 8,780	Samples missing.

8,718 *Nordegg Beds* 97 Ft.

Interval (Ft.)	
8,780 to 8,790	Shales, black; fissile and pyritic shales. Few calcareous shales, black.
8,790 to 8,810	Shales, black; few calcareous shales; shales with silty-sandy laminæ. Limestones, black, argillaceous; limestones, black, argillaceous, with insets of recrystallized calcite; black, phosphatic limestones and shales. Few siltstones and sandstones, clean or argillaceous. Chert, black, minor. Few shell fragments.

SCHOOLER CREEK GROUP

8,815 *Grey Beds* +588 Ft.

(Baldonnel Formation)

8,810 to 8,820	Cuttings, as for 8,790 to 8,810 ft. Siltstones, with very fine-grained sandstones, calcareous, dolomitic, and quartzitic. Limestones, variably dolomitic, and dolomites, silty. Shales and pyritic shales, black.
8,820 to 8,840	Shales and pyritic shales, black. Dolomites, light to brown coloured, microcrystalline and partly recrystallized; silty dolomites. Siltstones and sandstones, very fine grained, calcareous to dolomitic.
8,840 to 8,860	Bedded chert, mottled brown; abundant in upper part. Dolomites, light coloured; silty and calcareous dolomites. Limestones, brown, microcrystalline. Siltstones and sandstones, very fine grained. Shales and pyritic shales, black.
8,860 to 8,880	As for 8,840 to 8,860 ft.—but more dolomites.
8,880 to 8,920	Dolomites, grey, brown, and light coloured, fine to microcrystalline, silty and non-silty. Dolomites, light coloured, sub-oolitic. Dolomites, light coloured, granular, drusy. Few siltstones, lesser sandstones; dolomitic or calcareous. Chert, minor, at 8,910 ft. Shales and pyritic shales, black; few arenaceous shales.
8,920 to 8,950	Dolomites and shales, as for 8,880 to 8,920 ft. Dolomites, brown, grey, and dark grey, microcrystalline, commonly in platy cuttings. Dolomites, silty, at 8,930 ft.
8,950 to 8,990	Dolomites and argillaceous dolomites, brown, grey, and dark grey, microcrystalline. Few siltstones and very fine-grained sandstones, argillaceous and dolomitic. Shales and pyritic shales; dolomitic shales; arenaceous shales.
8,990 to 9,020	Shales and pyritic shales, black; dolomitic shales, dark grey. Dolomites, brown to dark grey, fine to microcrystalline, and argillaceous. Chert, minor.
9,020 to 9,030	Shales and dolomites, as for 8,990 to 9,020 ft. Limestones, grey to brown, fine to microcrystalline.
9,030 to 9,060	Dolomites, grey and light coloured, dense, or sub-oolitic. Dolomites, light coloured, drusy. Dolomites, grey and light coloured, silty. Dolomites, brown to dark grey, fine to microcrystalline. Dolomites, brown to dark grey, partly recrystallized. Shales. Cavings.
9,060 to 9,100	Dolomites, brown to dark grey, partly recrystallized. Dolomites, light to grey coloured, fine, crystalline, or granular, silty. Few siltstones and sandstones. Shales and pyritic shales, black.
9,100 to 9,140	Dolomites, light grey and grey, fine to microcrystalline. Shales, pyritic, and arenaceous shales, black. Chert, minor.
9,140 to 9,170	Dolomites, light coloured, microcrystalline, dense, or drusy, or silty. Siltstones, minor. Shales.
9,170 to 9,240	Dolomites, light coloured, fine to microcrystalline, dense, or drusy, or silty. Calcareous and silty dolomites, and siltstones with carbonate matrices. Dolomites, grey to brown, microcrystalline, at 9,200 to 9,240 ft. Chert and quartz at 9,210 ft. Shales.

(Charlie Lake Formation)

9,240 to 9,330	Dolomites, light coloured, dense, or sub-oolitic, or drusy. Dolomites, grey to brown, microcrystalline. Dolomites, brown and dark coloured, partly recrystallized. Dolomites, buff to brown, with anhydritic intergrowths. Anhydrite, white, fine crystalline, in lesser amounts. Siltstones, dolomitic to anhydritic, silty dolomites and anhydrites, in minor amounts. Shales.
9,330 to 9,400	Dolomites, as for 9,240 to 9,330 feet. Anhydrites, white, fine crystalline. Siltstones; dolomitic to anhydritic siltstones. Few sandstones in lower part.

(*Total depth:* 9,403 ft., driller; 9,393 ft., Schlumberger.)

NOTE—GEOGRAPHIC NAMES IN THE PINE VALLEY

Many geographic names in the Pine Valley differ from those of former and colloquial use.

Little Boulder Creek is 5 miles northeast of the West Pine Bridge, and the name is given on the National Topographic Series, Sheet 93O. The creek is also known as Marten (?Martin) Creek by many residents, and its distributaries on the valley plain of the Pine as Lillico and Marten Creeks. Little Boulder Creek is next to Big Boulder Creek on the west. The two should not be confused with Bowlder Creek (formerly spelt Boulder), which is 2 miles west of Commotion Creek in the eastern part of the area.

The old names "(the) Middle Forks" for Twidwell Bend, and "(the) Lower Forks" for the confluence of the Pine and Murray Rivers reported by Selwyn (1877), Dawson (1881), and Spieker (1922) are still in use. The Murray River was called the "East Pine"; the Sukunka River, the "Middle Pine"; and the Pine, the "West Pine (River)." Centurion Creek is also known as Wabi Creek. The name Table Mountain of Selwyn's report (1877), and which is still used locally, has been replaced by Mount Wartenbe. Little Prairie is now renamed Chetwynd.

The National Topographic Series, Sheet 93P, first edition, named Wabi Hill according to local custom. Since then, the Canadian Board on Geographic Names approved the name Mount Wabi for the hill, latitude 55 degrees 40 minutes north, longitude 121 degrees 35 minutes west, and 1.25 miles south of the Hart Highway.

Crassie and Pass Creek are discarded synonyms for Crassier Creek, and Falls Creek another for Falling Creek. Silver Sands Creek is sometimes known as Gold Creek, the old name. The bridge of the John Hart-Peace River Highway over the Pine River near the east foot of Solitude Mountain is called the West Pine Bridge.

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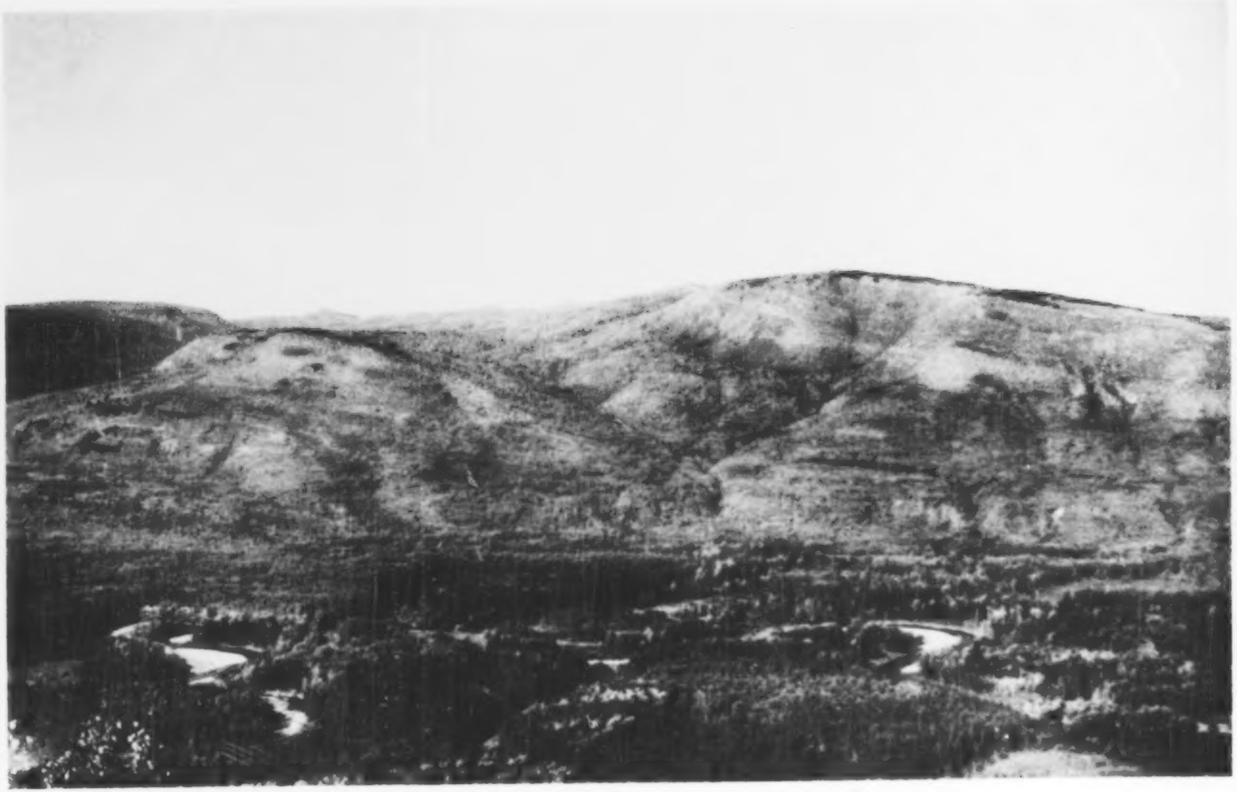

Plate I. Commotion Creek extreme left, Submarine Mountain left centre, Cruiser Mountain right centre, viewed across the Pine Valley from the high ground above Young Creek. The Commotion anticline on left (west), the Bissett syncline on the right, and the structural break between.

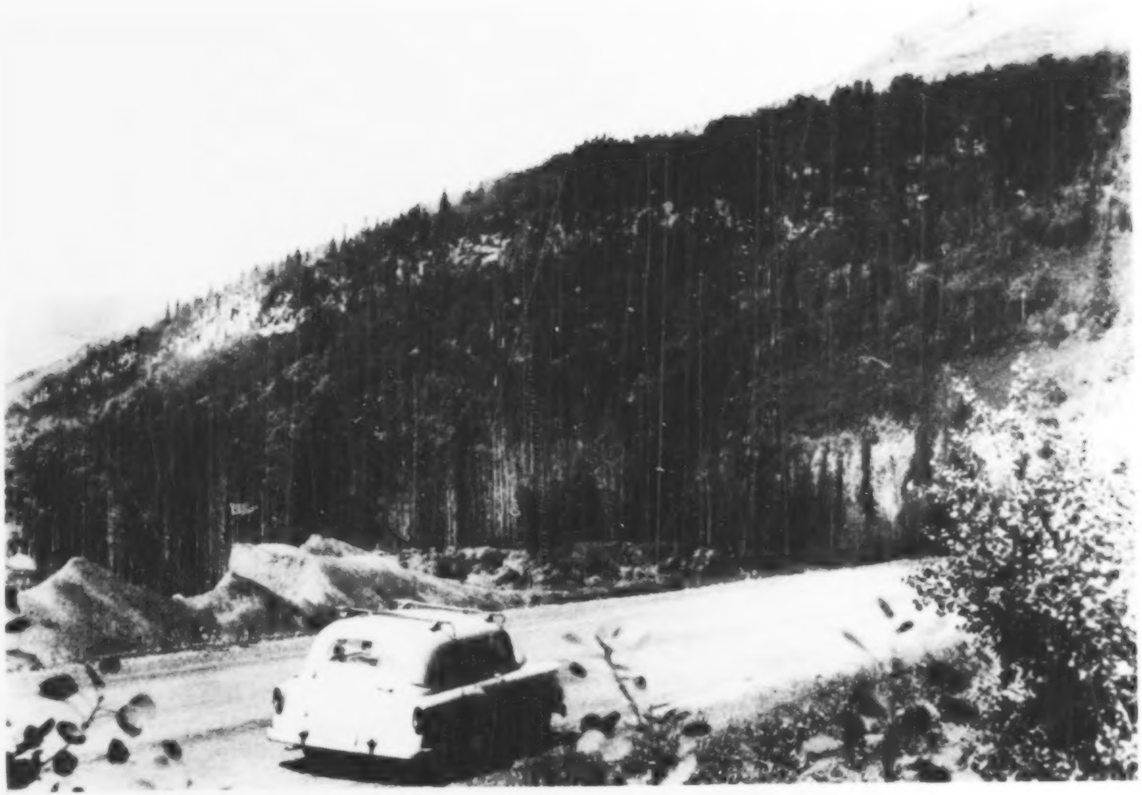

Plate II. The Commotion anticline at the John Hart-Peace River Highway (97). View westward from a point near the anticlinal crest. Conglomerate of the Commotion Formation (Member (iii)) exposed in the valley side; the Hasler Formation is hidden and the Goodrich Formation outcrops on the high ground at the right.

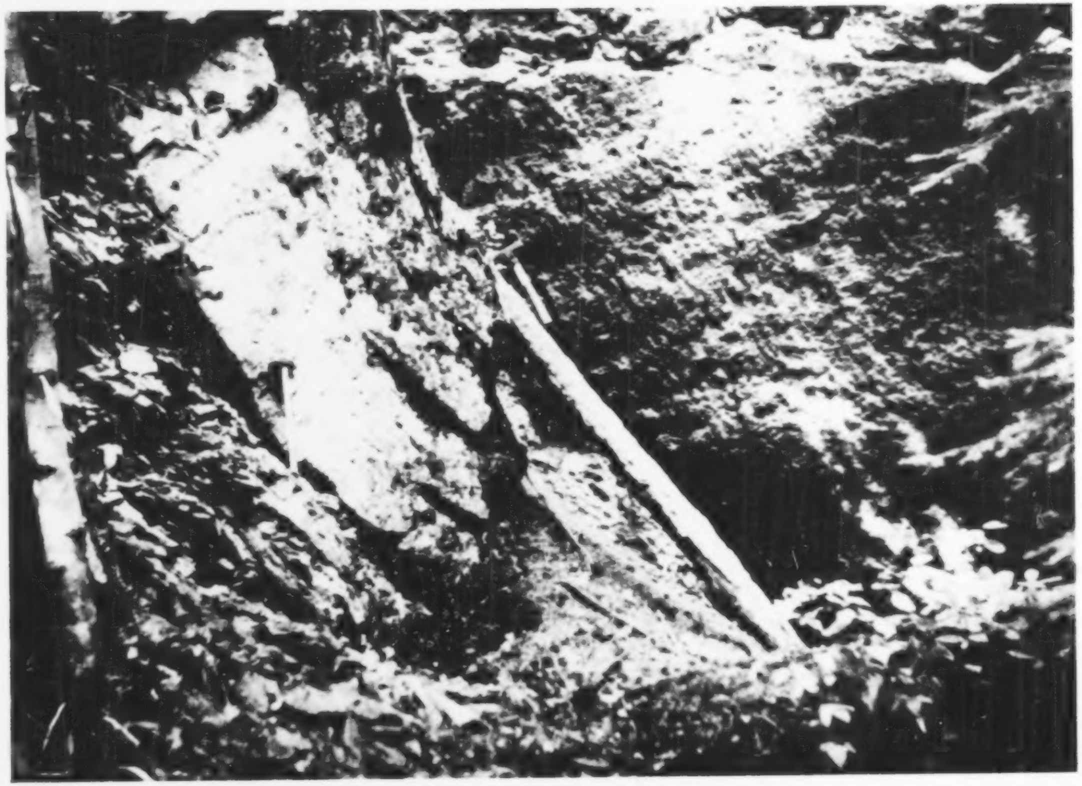

Plate III. The boundary of the Gething Formation and the basal conglomerate of the Moosebar Formation (left hammer) repeated by a thrust fault (along the line of burnt log marked by a second hammer). Fisher Creek, first tributary on west, upstream from the Hart Highway (97).

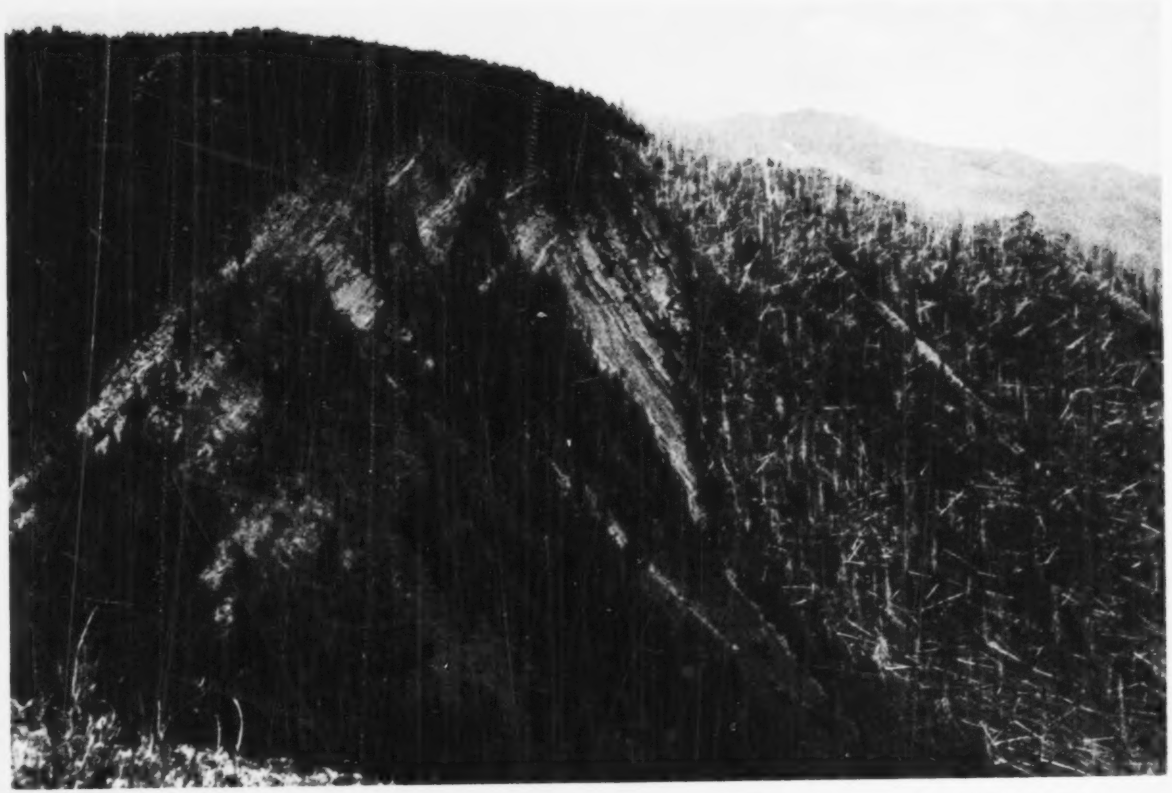

Plate IV. Angular fold in beds of the Crassier Group, 1 mile northeast of the eastern extremity of Bickford Lake.

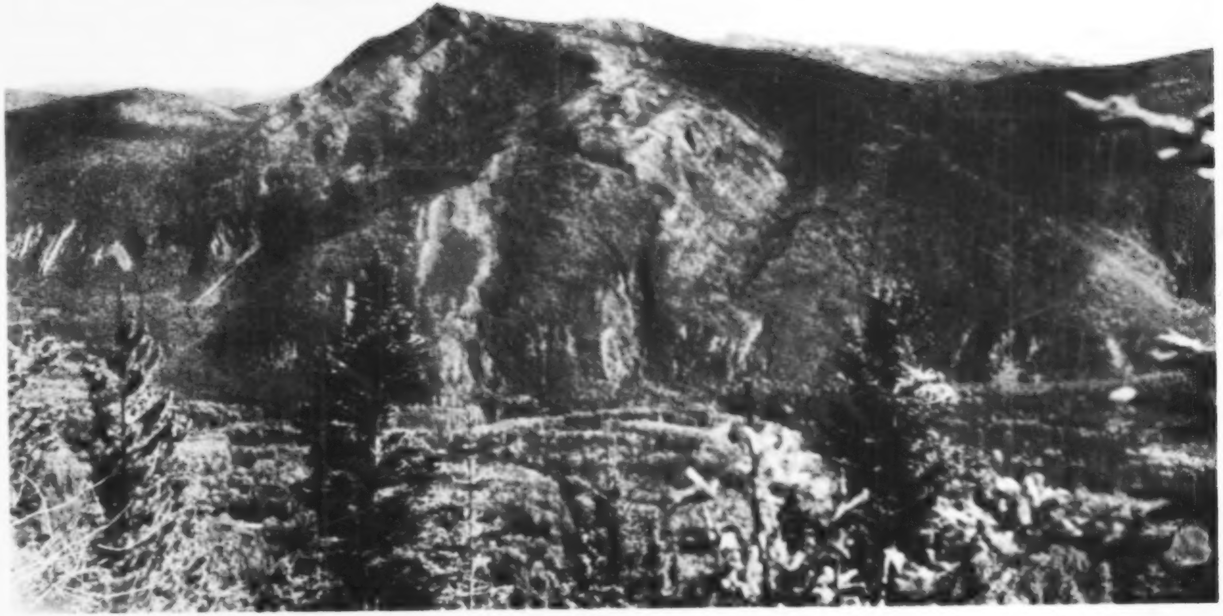

Plate V. The Bickford anticline on the south side of the Pine Valley. Axis of the Bickford anticline left (east), axis of the Coyote Creek syncline right of centre. Boundary of the Beaudette and Crassier Groups lies at dip slope very close to the summit peak.

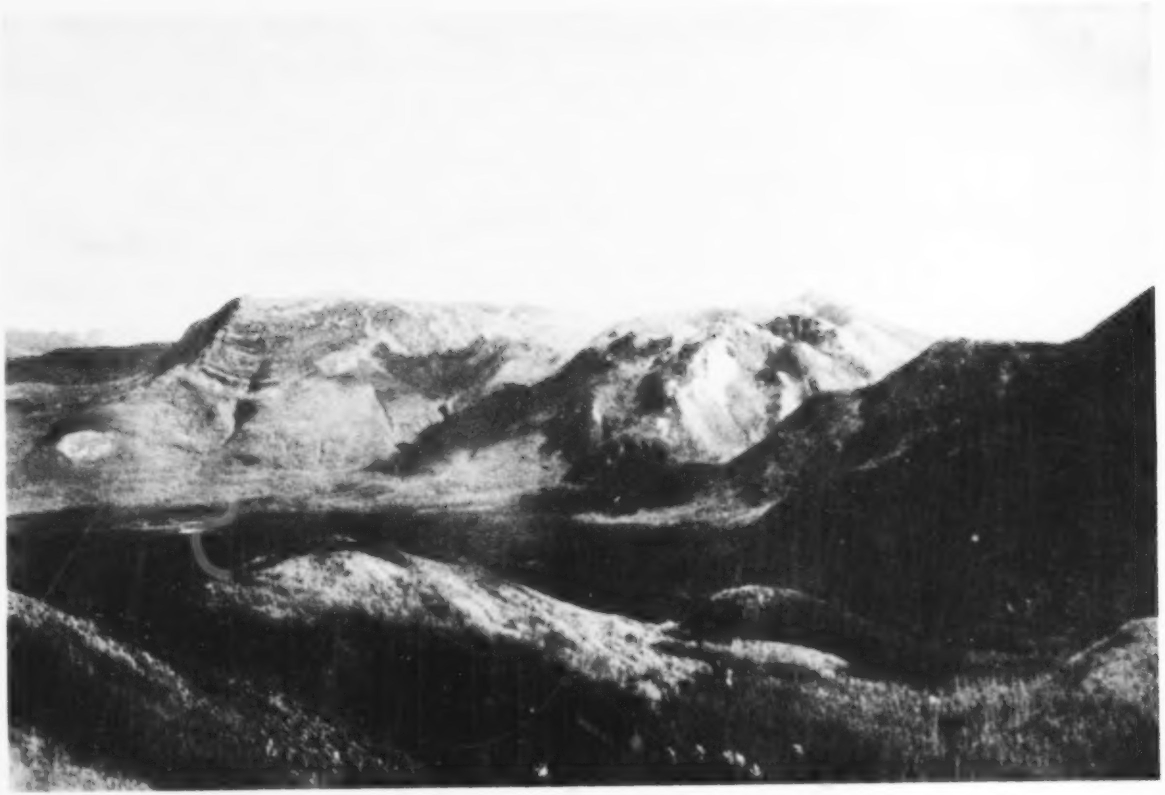

Plate VI. The north salient of Mount Le Moray, left centre. Right centre, Upper Palæozoic rocks thrust over Fernie and Triassic beds, in the Solitude thrust.